UNIVERSITY OF TASMANIA

GEOLOGY DEPARTMENT.

LOCAL AND REGIONAL EARTHQUAKES RECORDED BY THE TASMANIA SEISMIC NET.

by

I. D. RIPPER.

A Thesis submitted as part requirement for the Degree of Bachelor of Science with Honours, 1963.

CONTENTS.

ILLUSTRATIONS.

ABSTRACT.

Seismic research has been conducted in Tasmania since 1957, when the Fort Nelson station began operation in Hobart. Three additional stations, Tarraleah, Moorlands and Savannah, were installed in 1960 for the investigation of local seismicity.

Fifty-eight earthquakes occurred in Tasmania in the three year period from September, 1960 to December, 1963. The main activity was in the north west of Tasmania, and south of Macquarie Harbour, where the surface rock is mainly Precambrian and Cambrian sediments and metamorphics.

Seismic evidence suggests that the crust in central Tasmania is significantly thicker than the conventional 35 kilometers continental crust. High heat flow adds support to this view, although the Bouguer gravity anomaly is consistent with the 35 km value.

There is a regional distribution of earthquakes which give rise to T-phase recordings in Tasmania. This results from the effect of bathymetry on the oceanic Sofar channel. There is poor excitation of the T-phase where the continental margin slopes gently from 1000 to 4000 metres near the earthquake, and attenuation of the T-phase where the water depth is less

than 3000 - 4000 metres along the wave path. However, in some cases, considerable horizontal curvature around the South Tasmania Ridge occurred.

The average crustal thicknesses of the Tasman Sea and the South Tasmania Ridge were found to be approximately 12 km and 10 km respectively from Love wave dispersion.

1. INTRODUCTION.

1.1 Statement of Thesis

This thesis analyses the results of the first three years (1960-63) of seismic recording of local and regional earthquakes by the Tasmanian Seismic Net.

Section 2 deals briefly with the seismic net, and discusses its history, advantages, and the difficulties that have been encountered during its development.

Section 3 deals with the Tasmanian seismicity. Geological evidence of past seismicity, and earthquake reports during the last century precede a discussion of the Tasmanian seismic velocities and crustal structure, which are required to provide a model for the estimation of the local earthquake epicentres. Discussions of some major earthquakes recorded by the net, a summary of local seismicity during the past three years, and the earthquake distribution are presented. The consequences of the earthquake distribution and possible extensions of the seismic net to provide a better coverage of the island are also discussed.

The contribution made by the Tasmanian Seismic Net to the understanding of the nature and channel of the seismic T phase is presented in section 4. The regional distribution of earthquakes giving rise to T phases in Tasmania, and the factors which control the distribution are discussed. The main controlling factor is the bathymetry of the Tasman Sea, and some tentative conclusions are drawn concerning the effects of bathymetry on the Sofar channel. The seismic net is also used to examine the transmission across the island of the P waves derived from the T phase at the continental margin of Tasmania, and it is found that attenuation of the phase as it crosses the island and instrumentation differences result in the T phase records of the inland stations being generally weaker than those of Hobart.

Section 5 describes the use of surface waves, recorded mainly on the standard Sprengnethers installed in the TAU station of the World Wide Network, to determine average regional crustal structure around Tasmania. Since surface wave dispersion is dependent on crustal structure, it is being used extensively through out the world to obtain a more accurate knowledge of the crust and mantle.

The main achievements of the investigation are therefore (1) the mapping and discussion of current seismic activity in Tasmania. (2) The presentation and discussion of current evidence relating to the crustal structure of Tasmania. (3) The contribution of the Tasmanian seismic net to the knowledge of the T-wave and the Sofar channel, and (4) an estimation of the crustal thickness of the Tasman Sea and South Tasmania Ridge from surface wave dispersion.

1.2 Acknowledgements.

I wish to acknowledge the valuable assistance given by the staff and students of the University of Tasmania. In particular, Dr. Green, Miss Read, Miss Lobban, of the Geology Department, and Mr. Bartholomew, Mr. Ludbey and Mr. Bradshaw of the Engineering Department, have been most helpful in their advice and criticism.

2. THE SEISMIC NET.

2.1 Location.

The Tasmania seismic net began operating in September, 1960. It consists of four stations - Tasmanian University (TAU), Tarraleah (TRR), Moorlands (MOO), and Savannah (SAV). (fig. 1). Fort Nelson (FNT), situated in an underground concrete chamber at an old for on Mt. Nelson, was the original University station, but is now maintained for research only.

The seismic station positions were originally chosen to bracket Hydro Electric Commission installations in the central highlands of Tasmania, so that the regional seismic risk could be determined. All stations had to be within direct line of sight to Mt. Wellington or Mt. Barrow, which are repeater stations for the V.H.F. radio-telemeter system. They had to be situated on solid bedrock, and have electric power available. (University of Tas. Geol. Dept. Pub. No. 84).

The main disadvantage of the layout is that outlying parts of Tasmania, especially the northwest, are too far from the net to be efficiently monitored by it. The net itself encloses only one fifteenth of the area of the island.

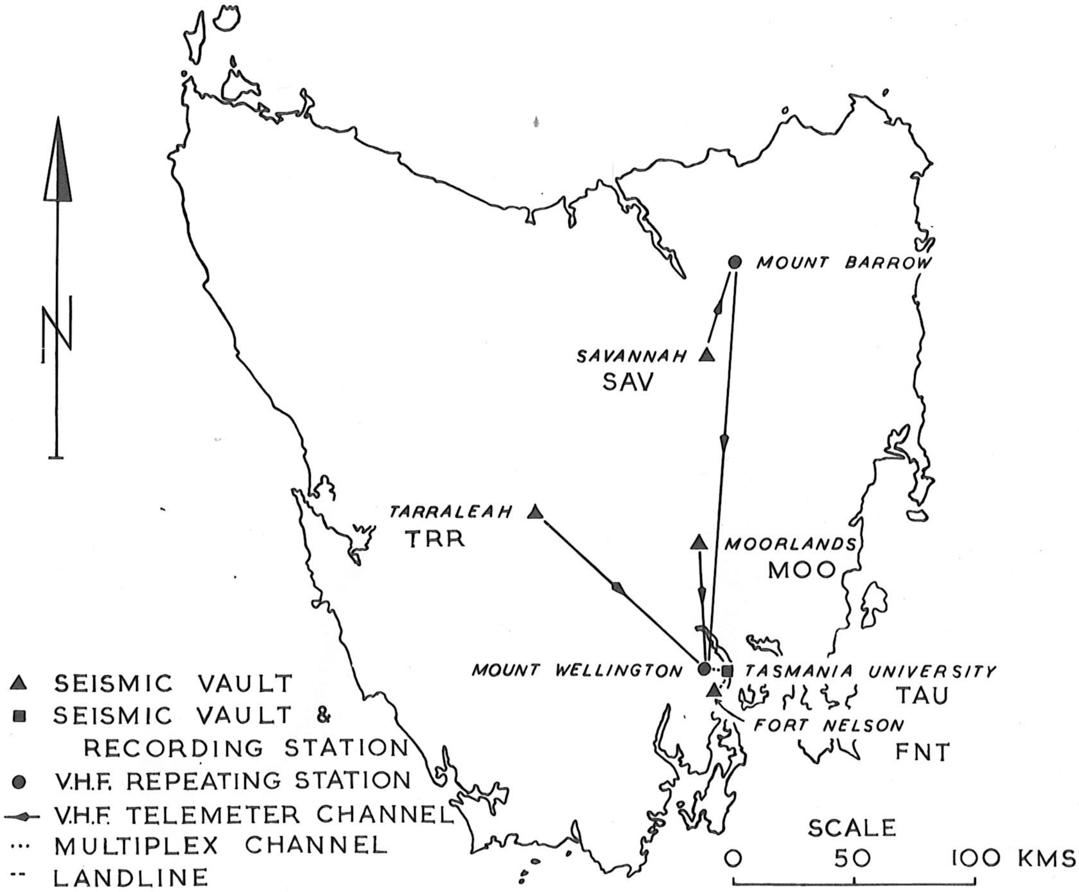

Fig.1. The Tasmania University Seismic Net. Tarraleah is now converted to land line transmission.

2.2 Instrumentation and Recording.

For local earthquake recording, each station is equipped with a Willmore short period (one second) vertical motion seismometer, which sits on a plinth of concrete poured onto solid bedrock. The foundation is basalt at Tarraleah, and dolerite at all other stations. The seismometer outputs are amplified and transmitted to the University by a V.H.F. radio-telemeter system or land-line, and fed to a pen recorder in the Geology building. The advantage of this system is that all stations appear side by side on the same paper strip, with the same timing system.

From 1957 until May, 1962 a photographically recording long period east west component Sprengnether Series H Galitzin type seismograph operated in Fort Nelson, and Willmore and Newstead - Watt seismometers operated intermittently. (Table I). Before September, 1960 the Willmore recorded on photographic paper, but with the inauguration of the seismic net, was changed to the six-pen recorder. The Newstead-Watt is still in the developmental stages. It is a horizontal motion variable reluctance A. C. seismometer with greater than usual sensitivity.

In May, 1962 a set of six seismometers

TABLE I. TASMANIAN SEISMIC STATION OPERATION, 1957 - 1963.

Fort Nelson (FNT)		
Willmore	Nov. 1957 - May, 1959	(Intermittently)
	Sep. 1960 - June, 1962	
Sprengnether	Nov. 1957 - April, 1962	
Newstead-Watt	April 1961 - June 1962	(Intermittently)
Tasmania University (TAU)		
Willmore	May 1962 - Dec. 1963	
Standard Set	May 1962 - Dec. 1963	
Newstead-Watt	Oct. 1963 - Dec. 1963	
Moorlands (MOO)		
Willmore	Sep. 1960 - Dec. 1963	
Tarraleah (TRR)		
Willmore	Sep. 1960 - Dec. 1963	
Savannah (SAV)		
Willmore	Oct. 1961 - March 1962	(Intermittently)
	August 1962 - July 1963	

were installed in the "TAU" vault by the United States Coast and Geodetic Survey, as part of the World Wide Standard Seismograph Network. Short period Benioff and long period Sprengnether seismometers with three matched components were installed in the vault situated on the hill behind the University. The Fort Nelson Willmore was transferred to the new vault, and the original Sprengnether ceased operation.

In May, 1963 a new ten-pen drum recorder was installed in the foyer of the Geology building, replacing the six-pen paper strip recorder. The timing system for the Standard Set installed by the U.S.C.G.S. was adapted for the local seismic net.

Since June, 1963 transmission from Tarraleah has been by a land line instead of the radio-telemeter system. This is designed to cut maintenance costs, and increase the efficiency of the station. If successful, the land line system will replace the radio-telemeter system.

The Willmore seismometers are not calibrated, so that relative or absolute amplitudes of the station signals cannot be determined. They have also been operating in an underdamped state, which tends to mask later arrivals.

The Hobart stations "FNT" and "TAU"

have consistently produced better earthquake recordings than the other stations, (Table II), mainly because they appear to be more sensitive to frequencies greater than one cycle per second. Moorlands is a particularly poor recorder of this frequency range. There is a high frequency cut-off at ten cycles per second in the radio-telemeter system, but this should not be detrimental to the performance of the net.

The poor high-frequency response of Moorlands is probably a characteristic of the site rather than the instrumentation. Some geological structures, such as a horizontal surface layer, sometimes accentuate certain frequencies because of resonance within the layer (Kanai, 1957). A thin layer of weathered granite has given rise to a greater proportion of higher frequencies than a newly exposed horizontal granite surface. (Kanai and Tanaka, 1961). Although high-frequency seismic waves are attenuated more than the lower frequencies, it appears that some factor of the Moorlands geology accentuates this high-frequency attenuation. The seismic vault at Moorlands is constructed upon the upper contact of a dolerite sill, which is several hundred feet thick. The Moorlands noise level is lower than the other stations, but this is misguiding since the

TABLE II. EFFICIENCY OF LOCAL EARTHQUAKE DETECTION BY SEISMIC NET STATIONS.

No. Recorded September, 1960 - December, 1963.

No. Recorded since SAV began operation, October, 1961; - 43.

	FNT - TAU	TRR	MOO	SAV
No. Recorded Usefully.	37	30	10	5
No. Poorly recorded.	14	13	23	2
No. Not recorded.	2	5	18	7
No. missed when station not operating.	5	10	7	27

corresponding local earthquake records are very poor, and quite often useless.

The noise level on Tarraleah is often increased by the operation of the Tarraleah Power Station about 1000 feet from the seismic vault, and by the flow of water down the flumes past the vault to the Power Station.

The Savannah seismograph has been the most troublesome, because of the additional repeating station on Mt. Barrow. In addition to normal maintenance problems, there has been considerable interference from Channel Six television transmission from Mt. Wellington. Channel Six operates in the frequency band 174 - 181 mc, and until recently, Mt. Barrow operated on 169.78 mc. Savannah is going onto land line transmission in January, 1964, and the problem will no longer exist.

Another cause of spasmodic recording has been the paper-strip jamming in the six-pen recorder. Although not a common occurence, it may have resulted on the loss of some local earthquake records. The paper speed (90 mm. per minute) was not constant, and the paper did not always pass evenly under the pens, so that the minute interval varied both along and across the paper. This resulted in a timing accuracy of no better than 0.5

seconds with respect to the minute marks. The timing marks on the record were corrected by ear, which was not as accurate as the present Stroboscope system on the Standard Set. Arrivals are now measured with an accuracy of one tenth of a second, Greenwich Mean Time.

Some stations are currently operating on two dynamic ranges so that strong earthquakes are recorded by a low gain pen which does not lose the detail of secondary arrivals.

3. TASMANIAN SEISMICITY.

3.1 Seismic History of Tasmania.

Evidence of past seismic activity is contained in the major pattern of horst and graben structures which date back to the Tertiary. The Derwent Graben, the Macquarie Harbour Graben, and the Cressy Graben are outstanding features of this system which trends mainly in a north-west direction. (Banks, in Spry and Banks (1962, page 241)). Current seismicity may well be the result of this activity still continuing.

More recent geological evidence of local activity, described in the Geology Department Publication Number 84 (1961), is the existence of Lake Edgar in south-west Tasmania. It is bordered by a fault which appears to be currently active, breaking through Quaternary sediments.

The first known earthquake to be felt in Tasmania was in January, 1924. It is described in a letter to Governor Sorell, but the location of the shock was not given. An investigation of Tasmanian seismicity since 1824 was carried out by Miss Read of the Geology Department, who compiled a detailed list from old newspaper reports and other correspondence. 2313 local earthquakes have been described between 1824 and 1962.

The majority of these, 2148 in number, were reported in a three year period, 1883 - 1886, and occured mostly in north-east Tasmania. An account of this sequence is given in the Publication Number 84 (1961). The figures presented in the publication have since been revised after further investigation.

The reported earthquake total (2313) is possibly an overstatement because it includes single reports which cannot be confirmed. Even so, reliable earthquake reports, that is, reports of an earthquake by more than one independant observer, show that Tasmania has experienced many large shocks, some with magnitude as high as VI, and it would not be unreasonable to expect strong earthquakes again in the future. Some local shocks were felt over the whole island, with maximum intensity VI on the Modified Mercalli Scale, (see Richter, 1958) which is equivalent to slight damage to normal buildings, such as fallen chimneys. There were three such shocks in 1884 and two in 1885. One shock, in July 1884, was reported with maximum intensity VI at Gould's Country, north-east Tasmania. It was felt throughout Tasmania, and reported in Victoria and New South Wales as well. The earthquake probably occured in the upper mantle to the north-east of Tasmania.

3.2 The New Years Day Earthquake, 1958.

The first local shocks recorded by instruments in Tasmania occurred on New Years Day, 1958, soon after Fort Nelson station began operation. The main earthquake was felt strongly at Queenstown (intensity V), Port Davey (IV), Blackwood Creek, near Great Lake, (IV), Getna (IV), and as far away as Burnie (III), Launceston (II), Hobart (II), and Swansea (II). (fig. 2).

This shock was too strong to give anything but the first arrival on the seismogram. However, two aftershocks gave both P and S wave arrivals on the Willmore seismograph at Fort Nelson. On the assumption that the aftershocks had the same epicentre as the main shock, the epicentral distance was approximately 140 kilometers, depending on the depth of focus. Reports received from the sparsely populated south-west were too few to determine the epicentre, and the best estimate was determined by combining the isoseismal information with the epicentral distance from Fort Nelson. (fig. 2). This placed it between the junction of the Gordon and Serpantine Rivers, and Mount King William, south of Lake St. Clair.

3.3 The Cygnet Earthquake, 1962.

The earthquake of August 21, 1962

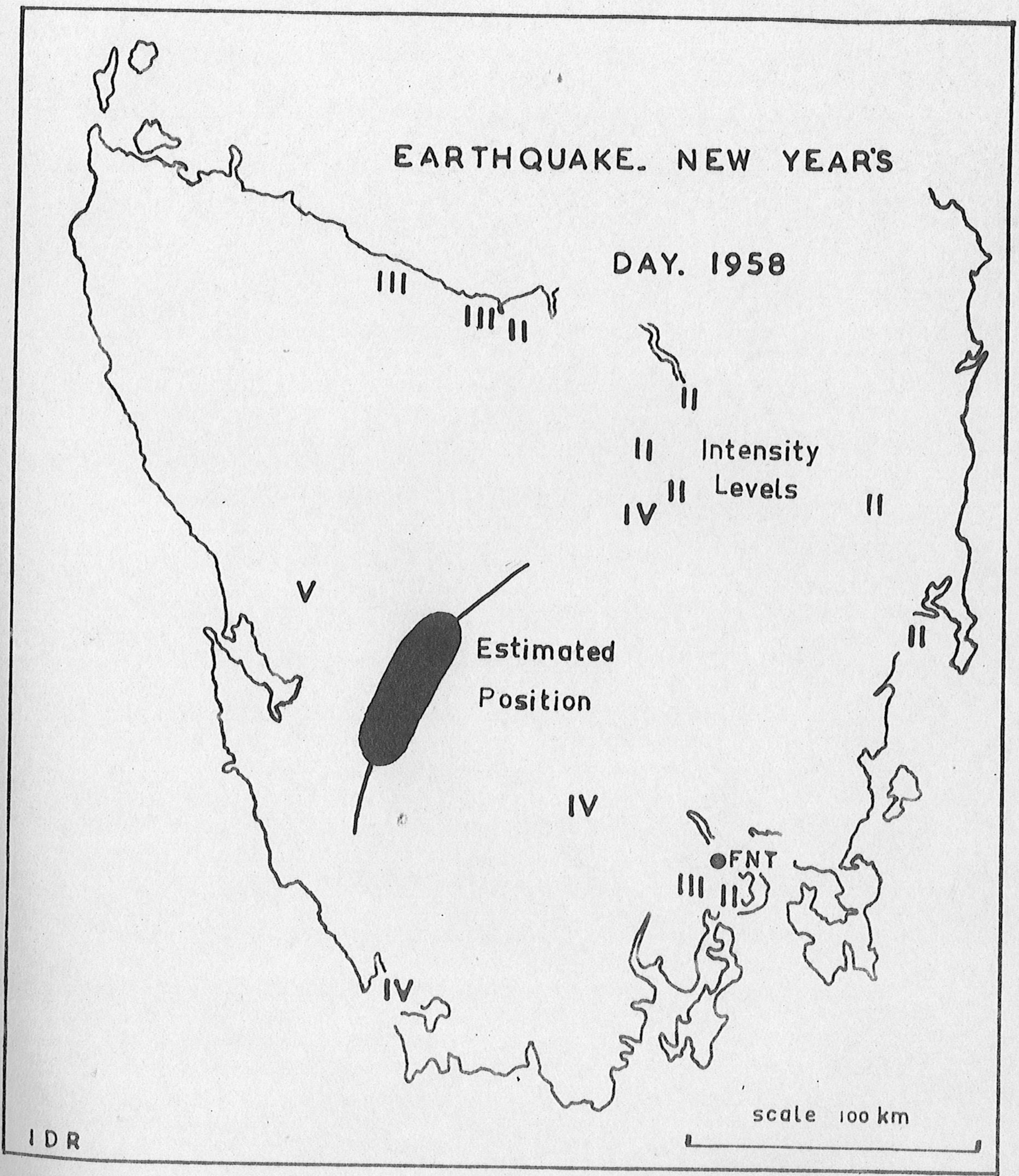

Fig. 2 Intensity reports and the estimated epicentre of the earthquake of 1st January, 1958.

(0632 a.m., August 22, local time) was felt in a small area around Cygnet, southern Tasmania. Questionnaires were distributed by the University, and based on this information, the intensity of the shock was III on the Modified Mercalli Scale. (fig. 3). The disturbance of a few seconds duration, awakened many people. The sensation was likened to thunder or a heavy truck. Rattling windows were widely noticed and there was one report, from Cygnet, of cracked plaster and damage to a concrete block.

The shock was felt only in a small area. (fig. 4). From the distribution of the reports, the epicentre was located to within 3 kilometers. This earthquake, although poorly recorded, provided the first opportunity to determine the crustal P velocity.

3.4 Crustal Velocity and Crustal Thickness of Tasmania.

Explosions are better for seismic crustal study than earthquakes because their origin time and position are accurately known. The epicentre of an earthquake must be known before it can be used for further study, and this usually requires a knowledge of the crust and the body wave velocities.

Since the Cygnet earthquake was felt in a small area, its epicentre is known to within about

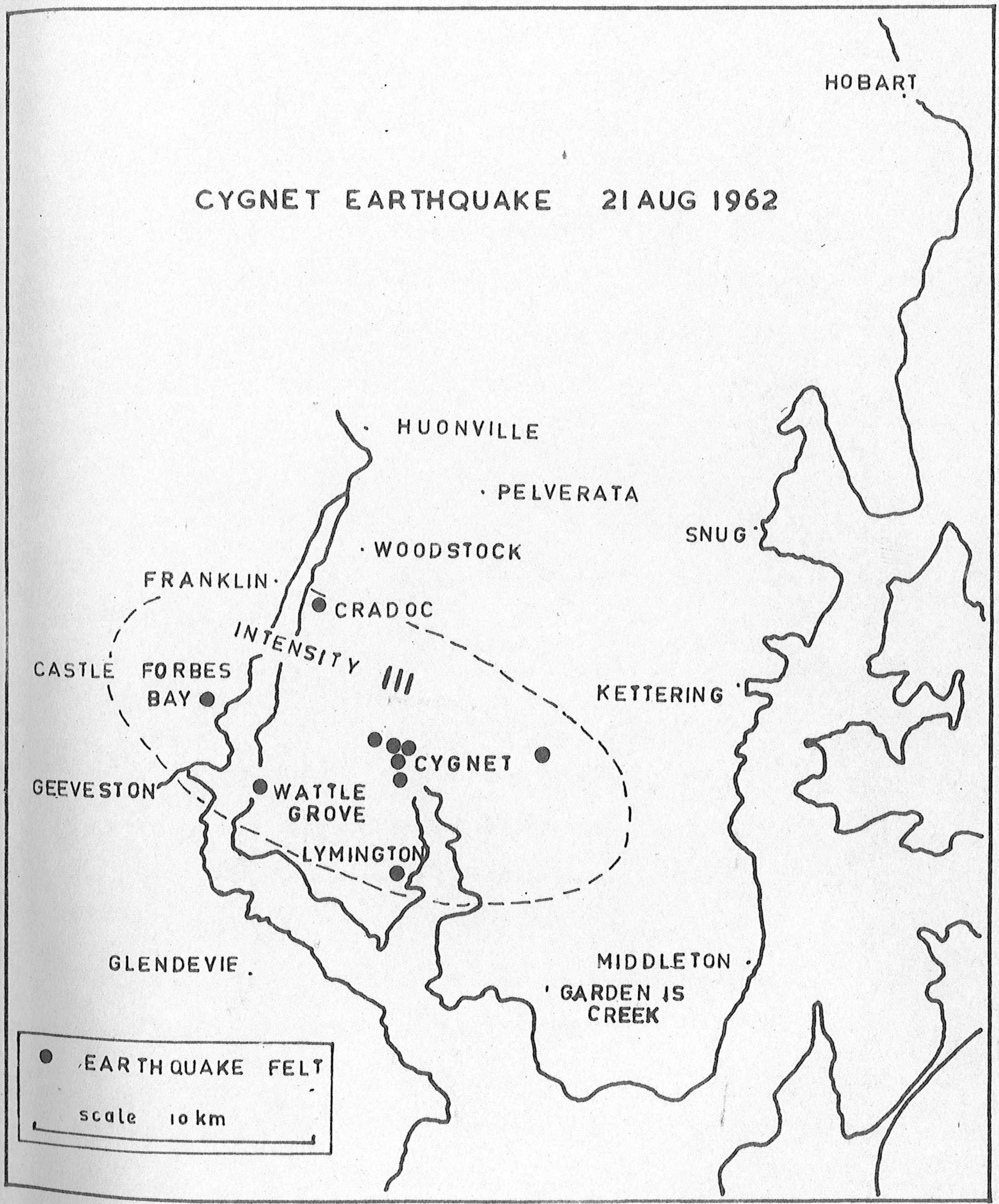

Fig. 3 The region in which the Cygnet earthquake, 21st August, 1962, was felt.

3 kilometers. Its depth is unknown. It was recorded with strong clear first arrivals on TAU and TRR, and with weak emergent first arrivals on MOO and SAV. Unfortunately the secondary (S) arrivals are emergent and indistinct.

The IBM 1620 computor at the Australian National University, Canberra, would not converge to a solution of origin time, epicentre and depth of focus. This is not surprising since only two arrivals are measured to within 0.2 seconds. All other arrivals may be a second out. The velocities used by the computer program (see Flynn, 1960) were determined for New South Wales by Doyle et al, 1959, from explosion studies. He obtained 6.03 and 3.61 km/sec for P and S respectively.

A method of successive approximations was applied to the P arrivals in an attempt to determine the crustal velocity. The four arrivals allow four equations involving the four unknowns, lattitude, longitude, origin time and the P wave velocity. An initial estimation of the value of each variable is made, and the equations are solved for the errors involved in the initial guesses. Squares and products of errors are deleted from the equations on the basis that they are insignificantly small. The equations are solved by a desk calculator, and the operation repeated until all errors are zero.

But this computation would not converge properly in the case of the Cygnet shock, or for any other local shock recorded by four stations and on the six-pen recorder because of the uncertainty of the measured arrivals. However, the method should be useful in the future for determinations of crustal P and S velocity from close earthquakes, without the aid of a computer.

The problem was then reduced to two simultaneous equations involving crustal velocity, and origin time. An epicentre at grid position (496, 686) was assumed, and the TAU and TRR P arrivals were used. A surface focus was assumed.

The Pg velocity obtained from this calculation was 5.9 ± 0.1 km/sec. The factors which introduce the inaccuracy are the uncertainty in the epicentre, which tends to reduce the calculated velocity, and the depth of focus, which tends to increase the velocity.

The S wave velocity was calculated from the assumed epicentre and the origin time calculated in the previous computation. It is 3.5 ± 0.1 km/sec.

The velocities are less than those for New South Wales obtained by Doyle et al (1959).

Information regarding crustal velocity and crustal thickness is obtained by measuring the wave front velocity across the net from local and regional shocks. The wave front is assumed to be plane, but it may be distorted by crustal irregularities. If the shock is close to the net, the curvature of the wave front will be significant. Since time is measured as differences between the stations, a small error in an arrival time may lead to a significant error in the velocity determination. For earthquakes in the north-west, the station triangles TRR, TAU, MOO, and TRR, FNT, MOO were used. In these cases, errors introduced by the assumption of a plane wave front are insignificant. North-west earthquakes 31, 47, 53 (Table III) gave rise to P wave velocities of 6.0, 5.7 and 5.9 km/sec respectively, the velocity of the Pg phase. The arrivals from the second and third shocks are weak and obscured by microseismic noise. The origin times calculated for each station are not in good agreement, but the errors are probably introduced in the S wave arrivals, which emerge from the P wave train. Therefore, the epicentres are not clearly defined.

If Pg arrived first at TRR and Pn at

TAU, the apparent velocity would be between 5.9 and 8.1 km/sec. If Pn arrived first at all stations, the apparent velocity would be 8.1 km/sec. But since the apparent velocity is 5.9 km/sec, Pg must have been the first arrival at all three stations.

The approximate minimum crustal thickness such that Pg arrive before Pn is given by

$$t = \frac{2H}{v\cos i} + \frac{d - 2H\tan i}{v'}$$

where

t	-	travel time to MOO
d	-	epicentral distance
H	-	crustal thickness
i	-	critical angle of refraction
v	-	Pg velocity
v'	-	Pn velocity

That is

$$H_{min} = \frac{t - \frac{d}{v'}}{\frac{2}{v\cos i} - \frac{2\tan i}{v'}}$$

where $\sin i = \frac{v}{v'} = \frac{5.9}{8.1}$

Hence, in kilometers

$$H_{min} = \frac{t - \frac{d}{8.86}}{0.2363}$$

This relation was applied to the MOO

arrival of shock 31, which produced clear Pg arrivals on all stations. MOO was chosen because it is halfway across the net from the north-west. The origin time of the earthquake was obtained from the TRR station which was closest to the earthquake.

The crustal thickness thus obtained was 60 km, which is abnormally thick. Although it seems from this that the crust of Tasmania beneath the Central Highlands is much thicker than normal, some factors may have contributed to make the obtained value too great. The Pn velocity has not been determined in Tasmania. If it is less 8.1 km/sec which was used in the calculation, the resulting calculated thickness will be too high. This suggests, then, that the Pn velocity is less than 8.1 km/sec. Also, if the Moho slopes down to the south east from the epicentre to the net, the Pn will arrive late, and so effect the calculation.

Jones (1962) showed that the Bouguer gravity anomaly in the Great Lake region, Central Tasmania is -30 to -40 milligal. This is consistent with a continental crust of normal thickness, 35 km. A thick crust is normally associated with a high negative Bouguer anomaly. (See fig. 4, from Wollard and Strange, 1962).

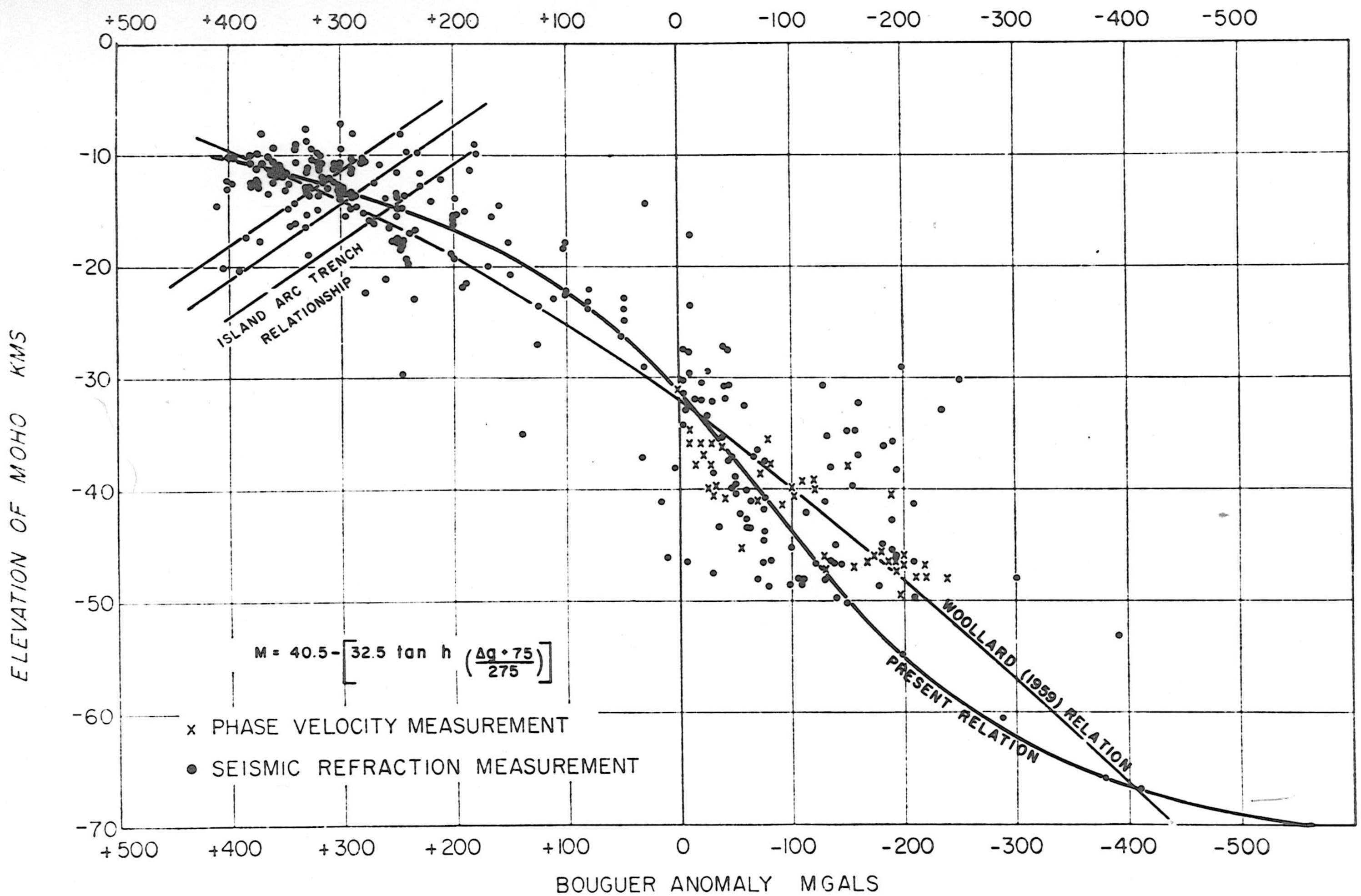

Fig.4. Relationship of elevation of M discontinuity to Bouguer anomaly. (After Woolard and Strange, 1962)

However, the higher density of the dolerite, 2.95 gm/cc compared to 2.8, the average for a crustal column, and 2.67 for upper crustal granitic layer would tend to make the Bouguer anomaly more positive. This would allow a thicker crustal column with a higher average density to give the measured Bouguer anomaly. Also, a denser crustal column, such as a crust with considerable dolerite in the upper portion, would have to be thicker than normal for isostatic equilubrium. But it seems impossible to explain the gravity readings of the Great Lake in terms of a crust as thick as that suggested by the seismic velocities. A crustal column with average density 2.95 gm/cc (the density of dolerite) would give a thickness of 55 km for isostatic equilubrium, but to assume the crust beneath central Tasmania consists of dolerite or rock of equivalent density is hardly valid.

A localised thickening of the crust beneath the Central Highlands, in the path of the seismic waves to the net, would not require as large a Bouguer anomaly as a uniformly thick crust. The Central Highlands rise above 1000 metres over an area 100 by 50 km, and in some places rise above 1600 metres. Another factor which is suggestive of a thick crust in Tasmania is the high heat flow, 2.0 μcal/cm^{2}sec, measured by Newstead and Beck (1953). Bullard and Griggs (1961) suggest that if the Moho represents a phase change and not a chemical change, then variations of Moho depth would be associated with corresponding variations in heat flow in the following manner -

Heat flow	1.1	1.2	1.6	2.0	2.4 μcal/cm^2
Moho depth	25	35	58	(71)	(80) km.

Their argument is that on the present evidence, the small variation of Moho depth for differing heat flows is negative evidence for a phase - change Moho. However, a high heat flow in Tasmania appears to be associated with a thicker than normal crust.

A study of body wave travel times failed to produce any definite information on crustal thickness. The earthquakes examined ranged in epicentral distance from 20° to 50°, and were recorded on the ten-pen recorder. No significant difference was found between the Jeffries - Bullen Seismological Tables (1958) and earthquake travel times to Hobart.

Some regional crustal thicknesses were obtained from surface wave dispersion, obtained by plotting group velocity against wave period. (Section 5). Surface wave dispersion measures crustal thickness on a regional scale and therefore has been of no value for the determination of the crustal thickness of Tasmania. But this may be possible in the future when identical long period Newstead - Watt seismometers are installed in all the seismic net stations. If they can be used to measure, with sufficient accuracy, the phase velocity of individual wave crests across the net, the crustal thickness could be obtained from the phase velocity dispersion curve. Hence, from the information currently available, our present knowledge of Tasmanian crustal structure can be summarised thus -

The Pg and Sg velocities are approximately 5.9 and 3.5 km/sec respectively.

The crust in Tasmania is significantly thicker than the normal 35 km continental crust. The maximum thickness is between 40 and 55 km, and is localised beneath the central Highlands.

The crust thins gradually to the south where the thickness is 10 - 14 km (section 5), and sharply to the east and west continental margins of Tasmania.

The average density of the crustal column is probably approximately 2.9 gm/cc, which is higher than the normal average of 2.8 gm/cc.

3.5 Location of Local Earthquake Epicentres.

The accurate location of local earthquakes is dependent upon,

(1) the crustal model,

(2) accurate identification of seismic phases, and

(3) accurate measurement of arrival times.

Most local earthquakes occur within 150 kilometers of the net, so that the first P and S arrivals are Pg and Sg, the direct waves through the upper crustal or "granitic" layer. For more distant shocks Sg may arrive after Sn, but the Sg carries the major portion of the seismic energy. This is the phase usually read as S, the Sn being completely obscured by the P wave train, and therefore missed by the interpreter. But the choice between Pg and Pn is critical, since a PnSg interpretation gives a much shorter epicentral distance than PgSg. For this study, a normal crustal thickness of 35 kilometers was assumed, with Pg and Sg velocities 5.9 and 3.5 km/sec respectively, and Pn and Sn velocities 8.1 and 4.7 km/sec respectively.

If an additional crustal layer exists in Tasmania - the "basaltic" or "gabbro" layer below the Conrad discontinuity - additional seismic phases P^* and S^* ~~will~~ would be propogated through this layer with velocities

6.5 and 3.7 km/sec respectively. These phases have not yet been identified in Tasmania.

In all cases except the Cygnet shock the epicentre was determined by calculating the distance of the earthquake from each station, and obtaining the intersection of the circles drawn with each station as centre and the epicentral distance as the radius of the corresponding circle. The circles should intersect at a point which is the earthquake epicentre. This method is sufficiently accurate for the present meagre knowledge of the local crustal structure.

3.6 <u>Earthquakes Recorded By The Seismic Net</u>.

In the period from October, 1957 to August, 1960, when only Fort Nelson was operating, only seven local earthquakes were recorded. The first three occured on New Years Day, 1958, and consisted of a main shock and two after shocks (see section 3.2). The long period Sprengnether seismograph installed in Fort Nelson was not intended for detection of local shocks, so that the small number of earthquakes recorded in this period is not surprising. A Willmore operated intermittently during 1958.

Except for the New Years Day shocks, the epicentres could not be determined as the earthquakes were recorded on only one station.

Inthe period from September, 1960 to December, 1963, 58 local shocks have been recorded by the seismic net. Their distribution in time is shown in figure 5. Of these, only 25 were recorded strongly enough for epicentre determination. The reasons for this are both seismic and instrumental. Some shocks were very weak, and are barely detectable above the microseismic level, or not present on the seismogram at all. Faulty recording at the time of an earthquake and intermittent recording have resulted in the loss of some earthquake records. The overall picture is given in Table II which shows that FNT and TAU have been the most reliable stations, followed by TRR. MOO and SAV have not produced many good local earthquake records.

Of the 25 shocks whose epicentres are shown in figure 6, 19 occurred in the western half of the island. Two shocks occurred off the east coast, and two off the south-east. Three shocks were recorded by two stations only and consequently have two alternate positions - i.e. the two intersections of the circles drawn about the two stations. Shock 15 occurred either near Port Davey or the Tamar River. The former position

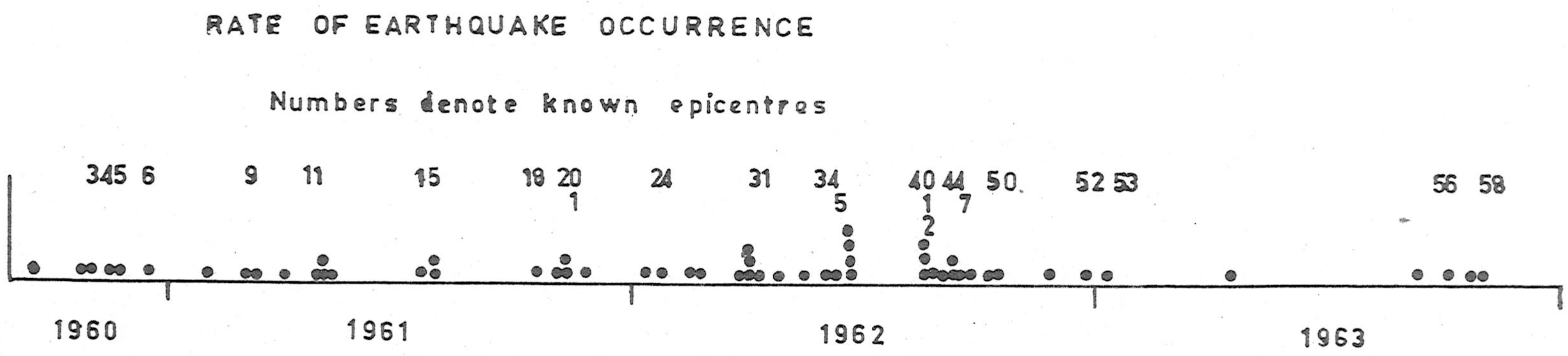

Fig. 5. The distribution in time of local earthquakes. The numbers refer to earthquakes shown in figure 6.

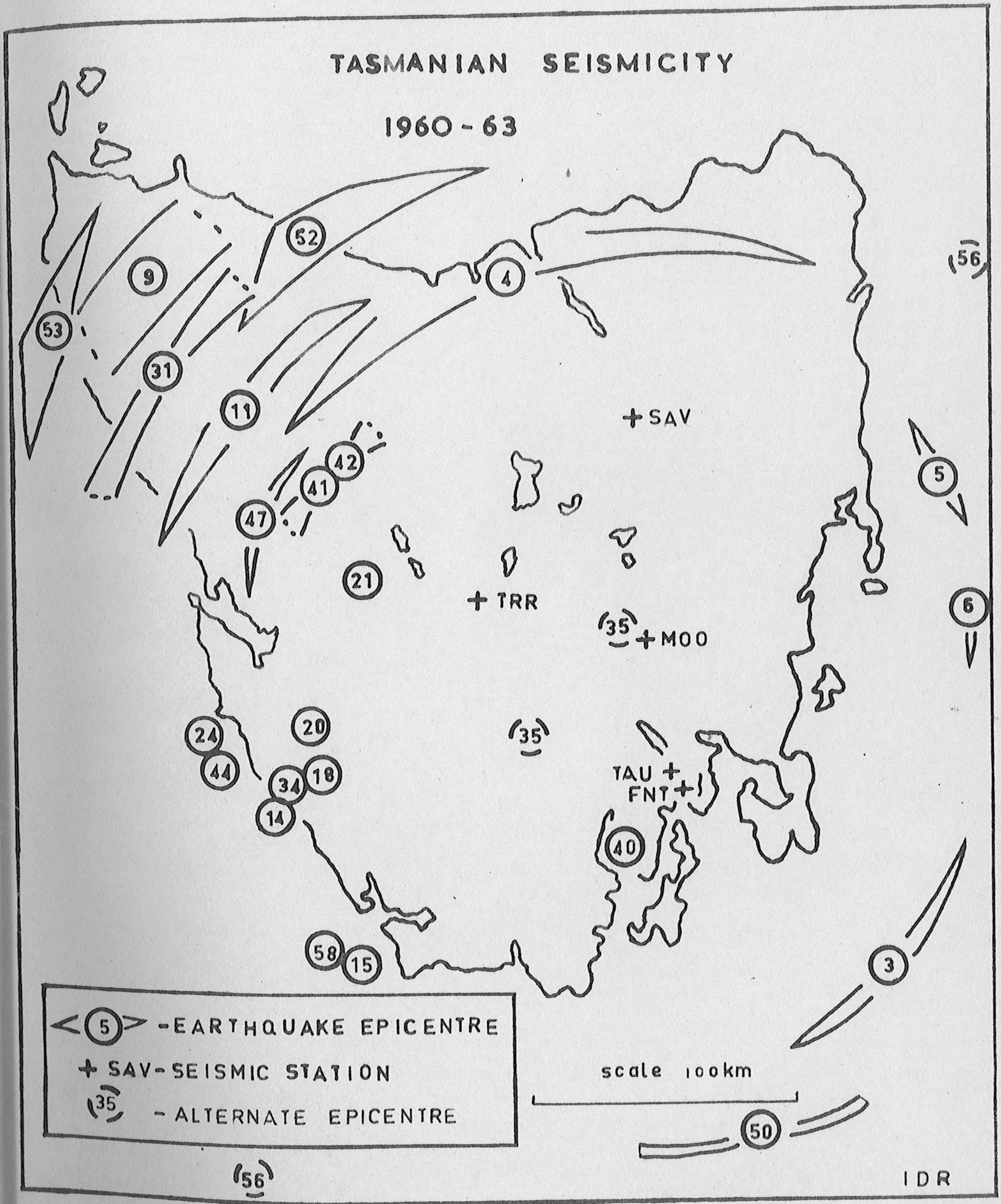

Fig. 6 The Distribution of Local earthquakes, Sept. 1960 to Dec. 1963. The dashed circles, in pairs, are alternate epicentres. Lines about the circles indicate the uncertainty of position. The numbers refer to Tavle III.

was chosen because it is in a seismically active region.

Epicentres in the south-west of the island are more accurately determined than those in the north-west. There are two main reasons for this.

1. The south-west shocks are within the epicentral distance for which Pg and Sg are the first arrivals. There is no uncertainty concerning the interpretation. North-west shocks are further from the net, and there is a choice between PgSg, PnSg and PnSn interpretations. The records are weak due to their greater epicentral distances.

2. In the south west, the circles which are drawn around the recording stations intersect at angles of about 50 degrees, and the epicentre is clearly defined. But in the north west, the angles of intersection are acute, and a small error in the radius of one circle has a large effect on the position of the earthquake. The uncertainty is in north-east to south-west direction. This is a disadvantage of the station locations of the seismic net for the recording of shocks in the north-west.

The epicentres of 33 shocks were not determined. Many of these were recorded only by the Hobart station, and probably originated to the south of Tasmania. Tarraleah was also the sole recorder of many shocks, which would have occurred in the extreme west of the island, or off the west coast.

Three earthquakes have been felt in Tasmania in the last six years, and recorded by the seismic stations. These are on 1/1/58, south of Queenstown, intensity VI; 21/8/62, Cygnet, intensity III; and 3/11/63, Port Davey, intensity V. Some northwest shocks may have been noticed, but they were not reported. There have been strong shocks in the north-west in the years preceding seismic recording in Tasmania.

3.7 Magnitude, Intensity and Depth of Local Shocks.

No direct local earthquake depth determinations have yet been possible. In all cases, the information has been insufficient. For a depth to be assigned with any certainty, the shock must occur within or very close to the seismic net, and must be recorded clearly by at least three stations, preferably by all four. These requirements have not been met.

Earthquake intensity is simply a measure of the effects of an earthquake, such as the amount of destruction of buildings. It is not a measure of the energy of the earthquake.

The first magnitude scale was proposed by Richter in 1935. Earthquake magnitude was defined as the logarithm (to the base 10) of the maximum amplitude (measured in microns) traced on a seismogram by a standard short period seismograph of period 0.8 sec, statical magnification 2800, damping coefficient 0.8, distant 100 kilometers from the epicentre. Magnitude is related to energy by

$$1.8\,M = \log_{10}\left(\frac{E}{E_0}\right),$$ where E_0 is

the energy of a shock of zero magnitude, taken as 10^{12} ergs, but uncertain by a factor of 10. (Gutenberg and Richter, 1942).

The Willmore seismometers of the seismic net are not calibrated, so that magnitudes cannot be calculated from the seismic records. An estimation of magnitude can be made from the approximate relationship between maximum intensity and earthquake magnitude for normal depth shocks, given in Richter (1958) page 353. From this relationship, the magnitude of most Tasmanian shocks is 1.5 and less. Bullen (1959) points out that the smallest felt earthquakes have magnitude 1.5. But many local shocks occured in uninhabited areas, and may not have been felt for this reason.

The Cygnet shock, maximum intensity III, was felt at a epicentral distance of 10 kilometers, and the corresponding magnitude is 2.5 approximately.

The New Years Day, 1958 shock was felt at a distance of 180 kilometers. The maximum intensity was VI, and the magnitude is set at 5, approximately.

Gzorsky (1962) gives the following relation between maximum intensity and magnitude,

$$I_{max} \simeq 1.5 M - \frac{h}{15}$$

where h is the focal depth. If we assume a depth of 15 kilometers, we get magnitudes 2.5 and 5 for the Cygnet and New Years Day shocks respectively.

From the energy relation, the energy of the smaller shocks was $10^{13} - 10^{15}$ ergs, the Cygnet earthquake, 10^{16} ergs, approximately. For comparison, atomic bomb blasts are about $10^{21} - 10^{23}$ ergs.

3.8 <u>The Port Davey Earthquake, 1963.</u>

On the evening of 3rd November, 1963 an earth tremor shook southern Tasmania. Unfortunately, the weather at the time was stormy, with blustery winds and rain, so that the shock passed unnoticed in many places. The P arrivals at the seismic stations saturated all the recording pens, except the Newstead-Watt, (TAU) pen. Thus TAU was the only station to record an S arrival. Also, three pens were recording TAU arrivals. The heavy drain on the power supply to the pen amplifiers corresponding to the P wave arrival at TAU caused cross feed to the other channels, giving a false first arrival at MOO.

The earthquake was located just off the south-west coast of Tasmania, at Port Davey (fig. 7).

The stormy weather had an adverse effect on the observation of the shock by the public. It was felt strongly at Port Davey, but not at Lake Pedder, 50 km away. It was felt strongly down the Huon Valley and on Bruny Island. There were only scattered reports from the west coast, Queenstown and Zeehan. It was not felt along the Lyell Highway, at Lake St. Clair or the surrounding Hydro installations. Approximate isoseismals are drawn in figure 7. The weather, or wind force at the moment of the shock may be a controlling factor of the isoseismal pattern, rather than conditions at the focus or geological structure.

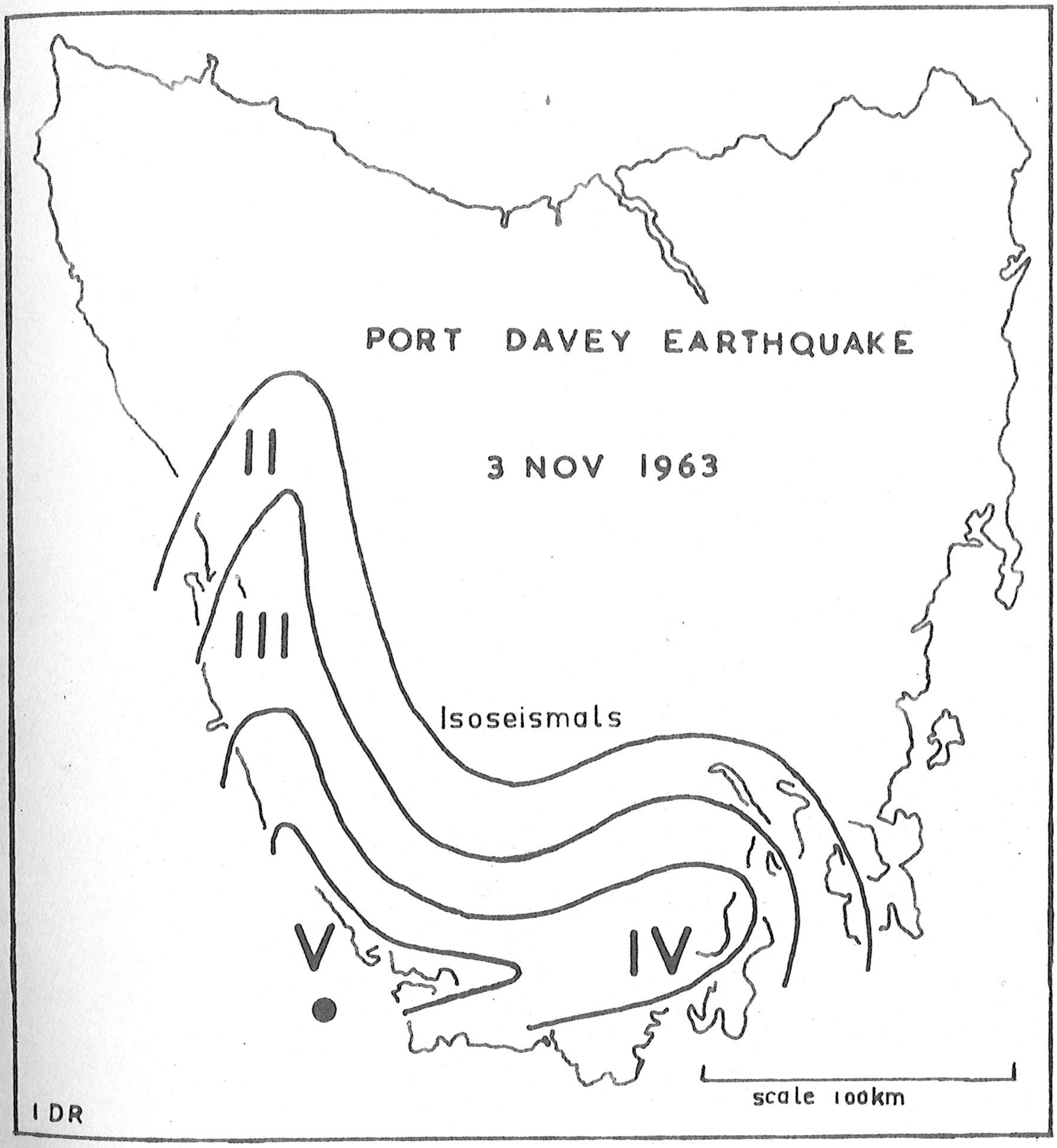

Fig. 7 The location and Isoseismals of the Port Davey Earthquake, 3rd November, 1963.

TABLE III. SUMMARY OF LOCAL SEISMICITY.

STATION		
1	-	FNT
2	-	TAU
3	-	TRR
4	-	MOO
5	-	SAV

No.	DATE	STATION	REMARKS
1.	23/ 9/60	1, 3, 4.	Local
2.	26/10/60	3.	Obscure on FNT, west Tasmania.
3.	2/11/60	1, 3.	South of Tasmania. (fig. 6)
4.	17/11/60	1, 4.	Poor MOO record, although closest station. Epicentre poorly defined in north Tasmania, at or between positions shown in figure 6.
5.	28/11/60	1, 3, 4.	Off east coast (fig. 6) Poor MOO record.
6.	21/12/60	1, 3, 4.	Off east coast (fig. 6).
7.	7/ 2/61	3.	Weak shock near TRR.
8.	3/ 3/61	1.	19 km from FNT, possibly quarry blast.
9.	14/ 3/61	1, 3, 4.	Poorly defined epicentre in

No.	DATE	STATIONS	REMARKS.
			north-west or off west coast (fig. 6) Poor MOO record.
10.	6/ 4/61	1.	Possibly local.
11.	26/ 4/61	1, 3, 4.	Epicentre in north-west (fig. 6) Poor MOO record.
12.	1/ 5/61	1.	Obscure on MOO, TRR. Prob. 75 km south of FNT.
13.	3/ 5/61	1, 3.	Weak Probably south-west Tasmania.
14.	9/ 5/61	1, 3, 4.	Weak on TRR, obscure on MOO. Approximate alternate positions, figure 6, late weak MOO reading suggests epicentre in south-west.
15.	19/ 7/61	3, 4.	Epicentre near Port Davey (fig. 6). Alternate position near Tamar River, but former position is in an active area; weak records,
16.	29/ 7/61	1.	Probably seismic.
17.	31/ 7/61	1.	Probably seismic.
18.	24/10/61	1, 3, 4.	Epicentre in south-west (fig. 6). MOO poor, but rejects alternate position in north.

No.	DATE	STATIONS	REMARKS.
19.	1/11/61	3, 4, 5.	Weak local shock.
20.	11/11/61	1, 3, 4, 5.	Strong south-west shock (fig. 6) south of Macquarie Harbour. P-wave saturated records, obscuring the S arrivals.
21.	15/11/61	1, 3, 4, 5.	Weak shock south-west of Lake St. Clair. (fig. 6) MOO, SAV poor records.
22.	29/11/61	3.	Weak shock near TRR.
23.	14/ 1/62	1.	Probably 41 km south of FNT.
24.	29/ 1/62	1, 3, 4.	Epicentre south of Macquarie Harbour (fig. 6) Alternate epicentre in north Tasmania eliminated by weak MOO record.
25.	21/ 2/62	1, 4.	Local south of Tasmania or in south-west.
26.	23/ 2/62	3.	Regional. Faint TRR record. 3 degrees from TRR.
27.	25/ 3/62	1.	Probably south of Hobart, epicentral distance from FNT, 100 km.
28. 29. 30.	5/ 5/62	1, 4.	3 shocks within 5 minutes. Probably related. Pens

No.	DATE	STATIONS	REMARKS.
			saturated by P arrivals. Best on Newstead-Watt. Epicentral distance to FNT, 175 km. MOO records weak. Probably south of Tasmania.
31.	12/ 4/62	1, 3, 4.	Epicentre in north-west Tasmania. (fig. 6) P velocity across net, 6.0 km/sec.
32.	25/ 4/62	1.	Distant 100 km from FNT, probably south of Tasmania.
33.	21/ 5/62	1.	Probably south or south-west Tasmania.
34.	1/ 6/62	1, 2, 3, 4.	Epicentre in south-west (fig. 6) Good records. First local shock recorded by standard set in TAU.
35.	7/ 6/62	1, 3.	Alternate positions Maydena or Apsley (fig. 6) or at greater depth between them.
36.	25/ 6/62	2, 3.	Probably local seismic.
37.	25/ 6/62	2, 3.	West or north-west, weak shock.

No.	DATE	STATIONS	REMARKS.
38.	25/ 6/62	2, 3.	Distant 70 km from TRR in west or north-west. The 3 shocks within 14 hours, probably related.
39.	27/ 6/62	1, 2, 3.	Weak on TRR strong on FNT, TAU distant 86 km from TAU. Probably south of Tasmania.
40.	21/ 8/62	2, 3, 4, 5.	Epicentre at Cygnet (fig. 6) Intensity III. Magnitude 2.5. Earthquake used to determine crustal velocities.
41.	22/ 8/62	2, 3.	Epicentre north-west of Lake St. Clair. (fig. 6) Not recorded by MOO, SAV.
42.	22/ 8/62	2, 3.	Aftershock, 3 hours after previous earthquake. Poor on MOO, not recorded by SAV.
43.	30/ 8/62	2, 3.	Weak shock.
44.	3/ 9/62	2, 3, 4.	Epicentre off west coast (fig. 6) Alternate position near Launceston eliminated as shock late on MOO, and not recorded by SAV.
45.	8/ 9/62	2, 3.	Records obscured by

No.	DATE	STATIONS	REMARKS.
			microseisms. Approximately 120 km from TRR in south or south-west.
46.	9/ 9/62	2, 3, 4.	Epicentre probably south-west. TAU masked by a previous record. Distant 180 km from TRR.
47.	21/ 9/62	2, 3, 4.	North-west of Lake St. Clair (fig. 6) Weak shock. P velocity across net 5.7 km/sec.
48.	23/ 9/62	2.	Weak shock. Distant 90 km, probably south of TAU, since not recorded by TRR, MOO.
49.	14/10/62	2.	Net not operating. Recorded on Standard Set only. Epicentral distance from TAU 165 km.
50.	18/10/62	2, 4.	Weak shock, south of Tasmania. (fig. 6).
51.	27/11/62	2, 3, 4.	Records obscured by noise. Distant 165 km from TRR, probably off west coast.
52.	26/12/62	2, 3, 4, 5.	Epicentre near north coast,

No.	DATE	STATIONS	REMARKS.
			(fig. 6) poorly defined, due to weak records.
53.	8/ 1/63	2, 3, 4, 5.	Epicentre poorly defined in far north-west or west. (fig. 6) MOO very poor. Velocity of P wave across net was 5.9 km/sec.
54.	19/ 4/63	3, 4, 5.	Weak shock, probably off west coast. Distant approximately 150 km from TRR.
55.	14/ 9/63	2, 3.	Good TAU record, but TRR weak, MOO obscure. Differences are mainly instrumental. Epicentral distance from TAU was 160 km.
56.	2/10/63	2, 3.	P arrivals weak. Alternate epicentres are approximately 100 km south-west of Port Davey, or approximately 40 km off north of east coast. (fig. 6)
57.	28/10/63	2, 3, 4.	Records are obscure. Possibly local.

No.	DATE	STATION	REMARKS.
58.	3/11/63	2, 3, 4.	Epicentre, Port Davey (fig. 6, 7). Intensity V. Magnitude 3.5.

3.9 Comparison With Geological Structure in Tasmania.

Current seismic activity in Tasmania is centred mainly in the western half of the island, with some activity south of Tasmania. (figs. 6, 8). The most strongly active area at present is south of Macquarie Harbour on the west coast. None of these earthquakes were felt, except the strong New Years Day shock in 1958 which occurred in the Gordon River area. The shock of 3rd November, 1963 which was felt in Hobart, occurred further south at Port Davey. The rock outcrop in this area is mainly Cambrian and Precambrian. Tertiary sediments are deposited in the Macquarie Graben, south of Macquarie Harbour.

The seismicity may be associated with movement along Tertiary faults of the Macquarie Graben structure.

The north-west is also seismically active. The area is populated, especially along the north coast and around Queenstown, so that although no shocks have been felt during the last six years, there is a small threat to public safety. A strong shock was felt in May, 1955 centred near Smithton. The maximum reported intensity was IV, on the Modified Mercalli Scale. Epicentres in this region have not been precisely

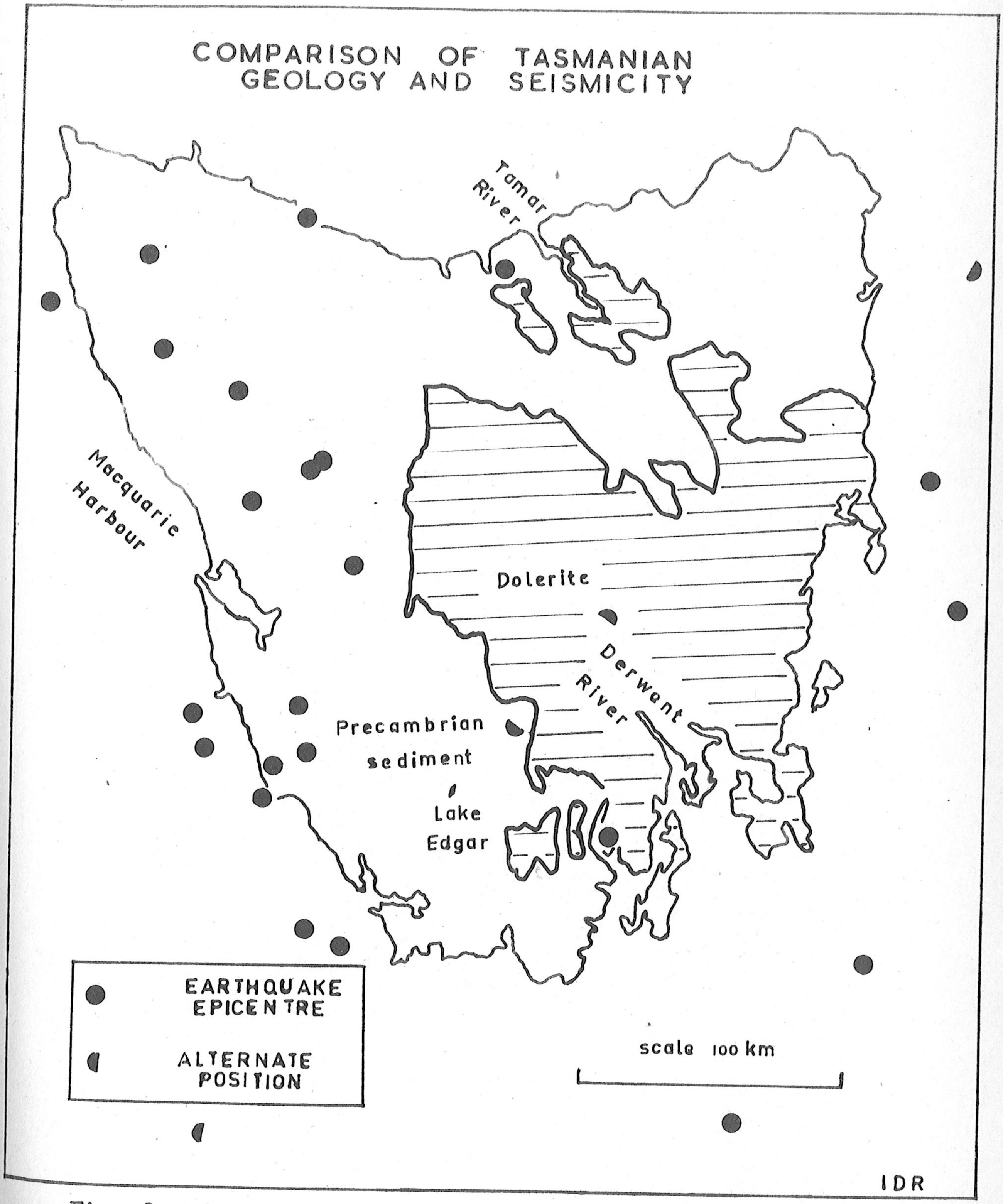

Fig. 8 Outstanding geological features of Tasmania - the Jurassic dolerite intrusion and the tertiary faulting - are shown for comparison with current seismicity.

determined, but placed within certain limits depending on the accuracy attained. The north-west consists mainly of Precambrian and Cambrian sediment.

A surprising feature of Tasmanian seismicity is that, since 1960, no earthquakes have occurred in central eastern Tasmania, although at least two have occurred off the east coast. Shocks have occurred up the western fringe the Jurassic dolerite, which dominates the centre and east of Tasmania, but no earthquakes have been positively detected within it (fig. 8). But the presence of strong dolerite faulting suggests that earthquakes do occur in the dolerite. In some cases, Jurassic faults are strongly expressed physiographically because of the resistant nature of the dolerite. (Banks, 1961). The resistant nature of the dolerite is possibly a reason for the low seismicity of eastern Tasmania in the period 1960-63. If the dolerite requires a higher stress for fracture than the Precambrian and Cambrian sediments of the west, then there will be fewer, but stronger earthquakes in the dolerite. The 1883-86 sequence of earthquakes in north-east Tasmania which included some shocks of magnitude as high as six may have been associated with the dolerite.

Thus, although earthquakes in the

centre and east are currently much less frequent than in western Tasmania, the seismic risk is significant because of the probability of high magnitude shocks, capable of considerable destruction. This is significant because of the high population and presence of large industries in central and east Tasmania, especially the Hydro-Electric Commission installations in the Central Highlands.

Alternatively, the west coast activity may be associated with flow in the upper mantle, localised below the west coast of Tasmania. The seismicity would result from the corresponding adjustment in the more rigid crust.

3.10 Future Seismic Net Developments.

The present seismic net (fig. 1) does not provide sufficient coverage for the island since the net is inadequate for earthquakes in the north-west. Another recording station in the north-west would increase the effectiveness of the net, and a site near Burnie would provide the best control for monitoring north-west shocks. Burnie is within 100 km of all the current north west activity, and the bearing to the area differs by 90° from the other net stations. This gives maximum control for the determination of epicentres. Burnie has electric power available and the Tertiary basalt would provide a solid foundation.

A transmission link from Burnie to Hobart would be longer than existing links. If land-line transmission proves successful for existing stations, it would be the better system for Burnie, as radio-transmission has not been as consistent.

If the recording apparatus were set up at Burnie, special care would have to be taken to ensure that time on the Burnie record was accurate to one tenth of a second with respect to the rest of the net.

A station in the south-west at Port Davey, would be a valuable addition to the net. But such an undertaking is not yet feasible. Savannah is well positioned to cover the north east, but so far, the station has been a poor recorder, (Table II) due to transmission difficulties associated with the radio link. This may be overcome when the radio transmission gives way to a land-line.

Moorlands is of little value in local earthquake recording (Table II) due to its characteristic of missing the high frequency component of the seismic signal.

The addition of Newstead-Watt seismometers operating on as high a frequency as possible about one cycle per second) will be invaluable in recording the S wave arrivals at the net stations. On the Willmores, these are often obscured by the P wave train. One Newstead-Watt in each station, oriented to record north-south ground motion, would significantly increase the efficiency of the net. The best orientation for a Newstead-Watt at Burnie would be east-west.

4. T PHASE RECORDED IN TASMANIA.

4.1 The T Phase.

One of the most striking earthquake phases recorded by the seismic net is the "T Phase", which was first reported by Linehan in 1940. It appears as a wave train of frequency 1 - 3 cycles per second on the short period Willmore seismograms about 10 - 12 minutes after the P wave, and lasts for several minutes. It is easily recognised by its characteristic appearance.

The T (third) phase is thought to be a combination of normal mode propogation through ocean bottom sediments with velocity 1.65 km/sec, and sound wave transmission through the oceanic Sofar channel, with a velocity of 1.49 km/sec (Northrop, 1962). The former is recorded as a very weak phase, and is often unrecognisable above the microseismic background, while the latter is the high energy arrival. At the continental margin, the energy is converted from the Sofar wave form to the P wave form, and it is this wave that is recorded by the land seismographs.

From December, 1960, until December, 1963, 90 T phases were recorded by the seismic net; 88 at Hobart, 30 at TRR, 24 at MOO and 7 at SAV. Hobart recorded many weak T phases which were attenuated before they reached the inland stations. This, combined with the fact that the Hobart stations have operated more consistently and more efficiently, explains the comparatively high number recorded by TAU and FNT.

The epicentres of 23 of the 90 shocks were given in the U.S.C.G.S. Preliminary Determination of Epicentre cards. 51 T phases were associated with clear P and S arrivals, and for the remaining 39, P and S were either very weak or absent. 32 T phase shocks were not reported on U.S.C.G.S. cards but had clear P and S arrivals from which the epicentral distance could be calculated. Table IV lists some of the major T phase earthquakes including those whose epicentres are given by the U.S.C.G.S.

4.2 Distribution of Earthquakes Which Produce T Phases In Tasmania.

Situated at the south east extremity of the Australian continent, Tasmania is ideally positioned to record T phases from New Zealand and Southern Ocean earthquakes. The regional earthquake distribution is given in figure 9, showing the earthquakes which produce T phases in Tasmania. A major seismic zone extends from Macquarie Island through New Zealand, the Kermadec Islands, Tonga Islands, Fiji, New Hebrides, curving north through the Solomon Islands and through New Guinea. South of Tasmania, the zone joins a minor seismic belt which passes

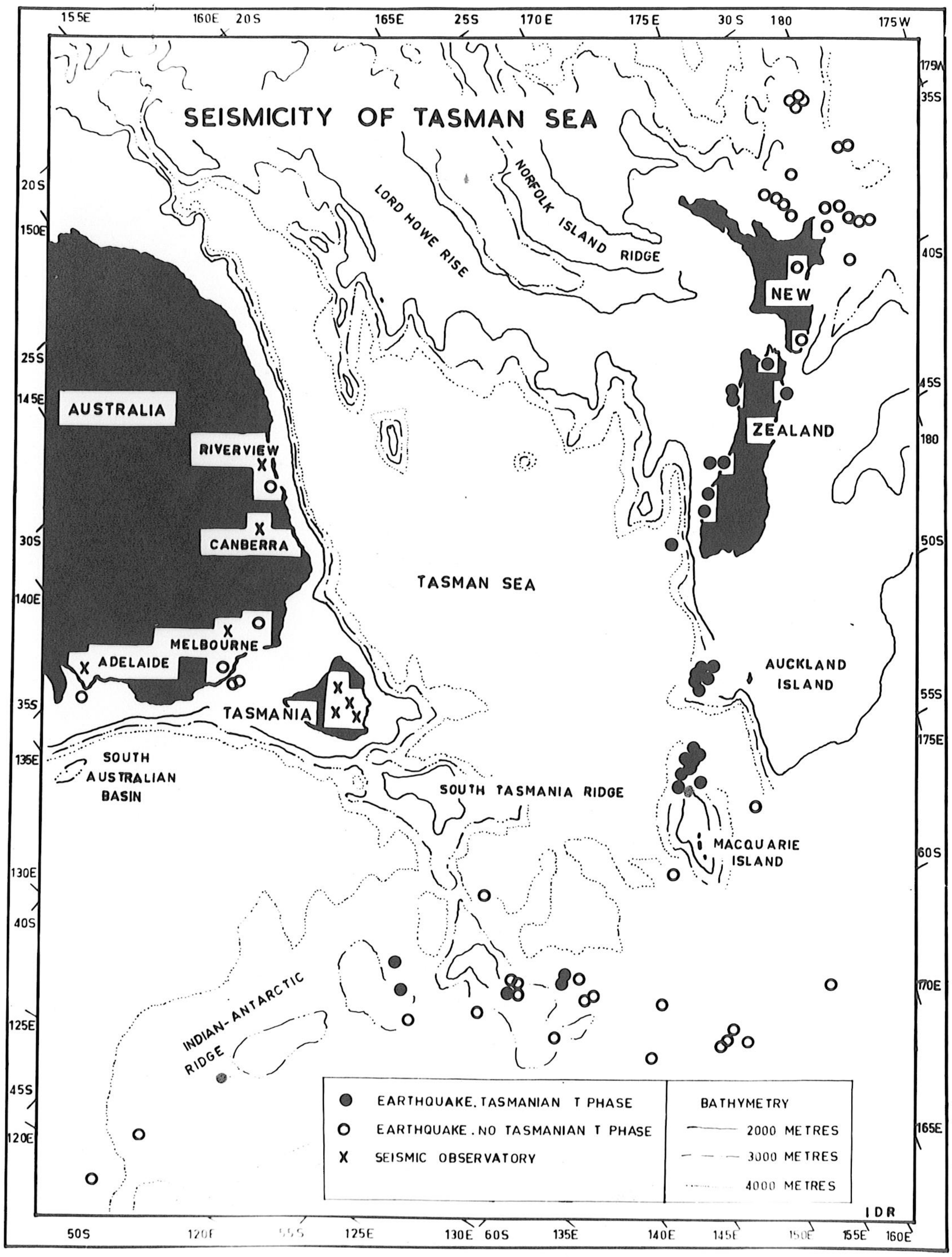

Fig.9. New Zealand and Macquarie Island earthquakes recorded in Tasmania, showing earthquakes which produced Tasmanian T-phase recordings, other Australian stations which record T-waves, and the bathymetry which controls the Sofar channel.

between Australia and Antarctic at latitude 50S - 60S°. Of these regions, earthquakes of the South Island, New Zealand, Aucklands Island and Macquarie Island give rise to strong T phases (fig. 9). Occasional T phases originate southwest of Tasmania, and three very weak T phases have originated due south of Tasmania. T phases have not been recorded from earthquakes further north than lattitude 41S, which separates the North and South Islands of New Zealand, or from earthquakes in the Southern Ocean, south of the Australian continent. Most of the T phase earthquakes which were associated with P and S arrivals in Tasmania, but whose epicentres were not reported by the U.S.C.G.S., probably occurred between Macquarie Island and New Zealand (fig. 9), because most of the calculated epicentral distances were about twelve degrees, agreeing with the distance from this major T phase source. (Example, Table IV, No. 31).

The T phases recorded without preceding P and S waves (example Table I.V, No. 35) may have resulted from phenomena such as submarine landslides off the New Zealand coast or around the margin of the Tasman Sea, and weak shallow earthquakes which injected large amounts of energy into the Sofar channel but whose P and S waves were too weak to be recorded in Tasmania.

New Zealand T waves have been reported in the seismic bulletins of observatories in Victoria and New South Wales. (Table I.V.) Burke - Gaffney (1954) first reported the New Zealand phase from the Riverview, Sydney seismic observatory. The Australian National University observatory at Canberra and the Department of National Development observatory at Toolangi, Melbourne, record T waves from New Zealand earthquakes. (fig. 9). The University of Adelaide observatory records T waves from south-west of Tasmania, and it has recorded at least one (No. 15) from Auckland Island.

Green (1962) first reported the presence of T phases on Tasmanian records, and he suggested that the T phase effect is due to New Zealand crustal structure. However, the problem involves not only the crustal structure and bathymetry near the epicentre, but also the bathymetry along the transmission path. The former, associated with the earthquake magnitude and depth, controls the energy injection into the channel. The latter controls the efficiency and extent of the Sofar propogation.

4.3 Bathymetry Control on The Sofar Channel.

The bathymetry of the Tasman Sea is shown in figure 9, and is taken from the Times Atlas of the World, 1958. The relation between bathymetry and

T phase earthquakes is very noticable.

The earthquakes which occur north of the 41°S lattitude through New Zealand do not give rise to T phases in Tasmania. The two reasons for this are, the long distance that the seismic energy must travel through the earth before it can be converted to T wave energy, and the gentle slope of the sea floor between the 1000 and 3000 metre isobaths, which is unsuitable for efficient excitation of the T phase in the Sofar channel. The first ten degrees of path between the earthquakes and Tasmania consist of a series of undersea ridges, and the water depth is less than 1500 metres. Therefore, the seismic energy must be transmitted as P and S waves for this distance before it can be transformed to T wave energy. But P and S waves spread out in three dimensions from the epicentre, and their amplitudes are proportional to $\frac{1}{r}$, where r is the distance from the epicentre (Ewing and Press, 1956). Thus the long transmission through the earth considerably reduces the energy available for conversion to T waves at edge of the Sofar channel.

In deep oceans, the channel is formed by the combination of temperature and pressure variation with water depth. The velocity minimum is at a depth of 1500 metres and the velocity increase below this depth is very gradual. (Ewing, Jardetzky, Press, 1957, page 336). If the water depth is only slightly greater than 1500 metres, the velocity increase which forms the bottom refactor of the Sofar Channel is absent and energy is lost into the sea bed. The effects of a shallow sea bed, such as variable salinity and water currents of different temberatures giving a variable temperature distribution would disturb the sound velocity distribution near the surface. (Officer, 1958). If such variations extend to about 1500 metres, they would distort the Sofar channel itself. Hence, for a shallow gently sloping floor, the Sofar channel would not exist until the depth was much deeper than 1500 metres, and conditions would be unfavourable for energy injection into the channel.

These two factors explain the lack of Tasmanian records of T phases from north of New Zealand. No T waves have been recorded from any earthquakes north of New Zealand or north-west of Australia, and the presence of the Lord Howe Rise and other submarine ridges in the north Tasman Sea is sufficient to explain this.

The sea floor off the west coast of the South Island drops sharply to a depth of 4000 metres, and strong T phases from this region are recorded in Tasmania. Therefore, a steep continental slope must provide good coupling between continent and ocean for energy injection into the Sofar channel, and the Sofar Channel must terminate sharply against the continental margin. Two shocks in the north and north east of the South Island (Table IV Nos. 27, 32) produced very weak, barely discernable T records at TAU. Canberra reported a T wave from a shock in the southern tip of the North Island, at 41.2S, 175.7E, on 27/12/1961. There was no trace of this T wave on TAU, but the faint record may have been obscured by the high microseismic level at the time.

Further south, strong T phases originate from all shocks between Macquarie Island and New Zealand (fig. 9). Two main centres exist. The first is near Auckland Island, and the second 3 degrees north of Macquarie Island. No submarine ridges exist between Tasmania and these epicentres. The path is across the south Tasman Sea, where the water depth exceeds 4000 metres. Shock number 15 (Table IV) from this region was recorded in Adelaide. The T wave was recorded very strongly in Tasmania and the body wave train was strong enough to cross the Australian continent as far as Adelaide. The arrival was several minutes too early to be a T-wave which had travelled around the south of Tasmania.

An earthquake at 54.7S, 162.9E, east of Macquarie Island, on 11/1/1961 did not produce a T phase at Hobart, but the direct path to Tasmania crossed the 2000 metre isobath around Macquarie Island, which suggests that the Sofar channel is extinguished when the ocean bottom rises to this level. Another earthquake, five hours later, about 300 km north-west, was clear of this isobath and produced a T phase. (Table IV, No. 9). Canberra recorded a T wave from both shocks.

The ocean-sediment phase first described by Northrop (1962) is often present on records of earthquakes from the South Island of New Zealand, Auckland Island and Macquarie Island regions. It arrives about a minute before the strong Sofar signal as a very weak phase, hardly noticable above the microseismic background.

Two strong T phases were recorded from position (51S, 139E), south-east of Tasmania. They occurred between the 3000 and 4000 metre isobaths on the northern side of the Indian-Antarctic Ridge. The direct path touches the 3000 metre isobath south of Tasmania.

Many earthquakes originate in a region south of Tasmania, west of Macquarie Island, (fig. 9) and in general, do not give rise to Tasmanian T phase recordings.

This is to be expected because the channel is directly barred by the South Tasmania Ridge, which rises above the 1000 metre isobath. The path of one shock from 54.9S, 156.3E, on 5/7/1962 is along the 3000 metre isobath for about 500 km, but no T is recorded. This suggests that the T wave energy in the Sofar channel is gradually attenuated in a water layer of only 3000 metres.

There are, however, three exceptions in this region. Three earthquakes, (Table IV, Nos. 22, 39, 40) produced very faint T wave records. The shocks are masked by the South Tasmania Ridge, over which the Sofar channel cannot exist. Neighbouring shocks show no sign of a T wave.

The T arrivals are over a minute too late to be normal mode propogation through sedimentary sludge. The T waves from shocks 22 and 40 are also too strong. A cold water layer sandwiched between two warm layers may provide a freak T-wave channel over the ridge if such a condition existed at the time of the shock. But this explanation is unlikely.

The magnitudes of shocks 39 and 40 were $5\frac{1}{4}$ and $6\frac{1}{4}$, as reported in U.S.C.G.S. data cards. Magnitudes of other shocks in the area have not been given, which may indicate weaker earthquakes in the cases where T was not recorded in Tasmania. Sykes (1963) has shown that most Southern Ocean earthquakes have shallow focii, less than about 50 km. Therefore it seems that the T waves generated by shocks 22, 39 and 40 were stronger than from other shocks in the same area. The Tasmanian recordings of these T-phases is evidence for the diffraction or horizontal curvature of T-waves around an obstruction such as a land mass or submarine ridge. The T waves were recorded in Adelaide, so that the diffraction is probably around the western side of the ridge.

Shocks without Tasmanian recordings of T waves occurred at 48.9S, 122.8E on 16/8/1963 and at 48.2S, 119.4E on 23/5/1962 (fig. 9). The epicentres are on the Indian-Antarctic Ridge, and about 10 degrees of the direct path to Tasmania (half the path length) are at a depth between the 3000 and 4000 isobaths. A T-wave was recorded at Adelaide from the former shock, which shows that the absence of T in Tasmania is a result of attenuation in the Sofar channel, and not lack of excitation at the source. A closer shock on the Indian-Australian Ridge (No.42) produced a T-phase in Tasmania. Five degrees of path is between the 3000 and 4000 metres isobaths. This suggests that a water depth less than 3000 metres will gradually attenuate the T-phase over long distances.

The emergent nature of the T wave trains recorded in Tasmania is probably a result of injection of seismic energy into the Sofar channel over an appreciable area of the Tasman Sea, and for a significant period of time, so that the T wave gradually builds up. (In contrast to earthquake T phases, Northrop (1962) showed that T phases from explosions which occurred and were recorded within the Sofar channel begin and end very sharply on the seismic record.) The transfer of T wave energy from the Sofar channel to the continental crust along an irregular margin would also decrease the sharpness of the T arrival.

Reflections of the T-phase from submarine ridges and land masses, such as the South Tasmania Ridge and New Zealand, could contribute to the emergent nature and long duration of Tasmanian T-phase recordings. The variation of amplitude of a T phase from an earthquake (Table IV, No. 42), particularly noticable on the strong motion horizontal N-S Newstead-Watt seismogram, may be explained in this way. There are five clear peaks on the record. However, the formation of more than one phase (such as P and S) at the Tasmanian margin may also be involved in the amplitude variation of this T-phase. (fig. 10).

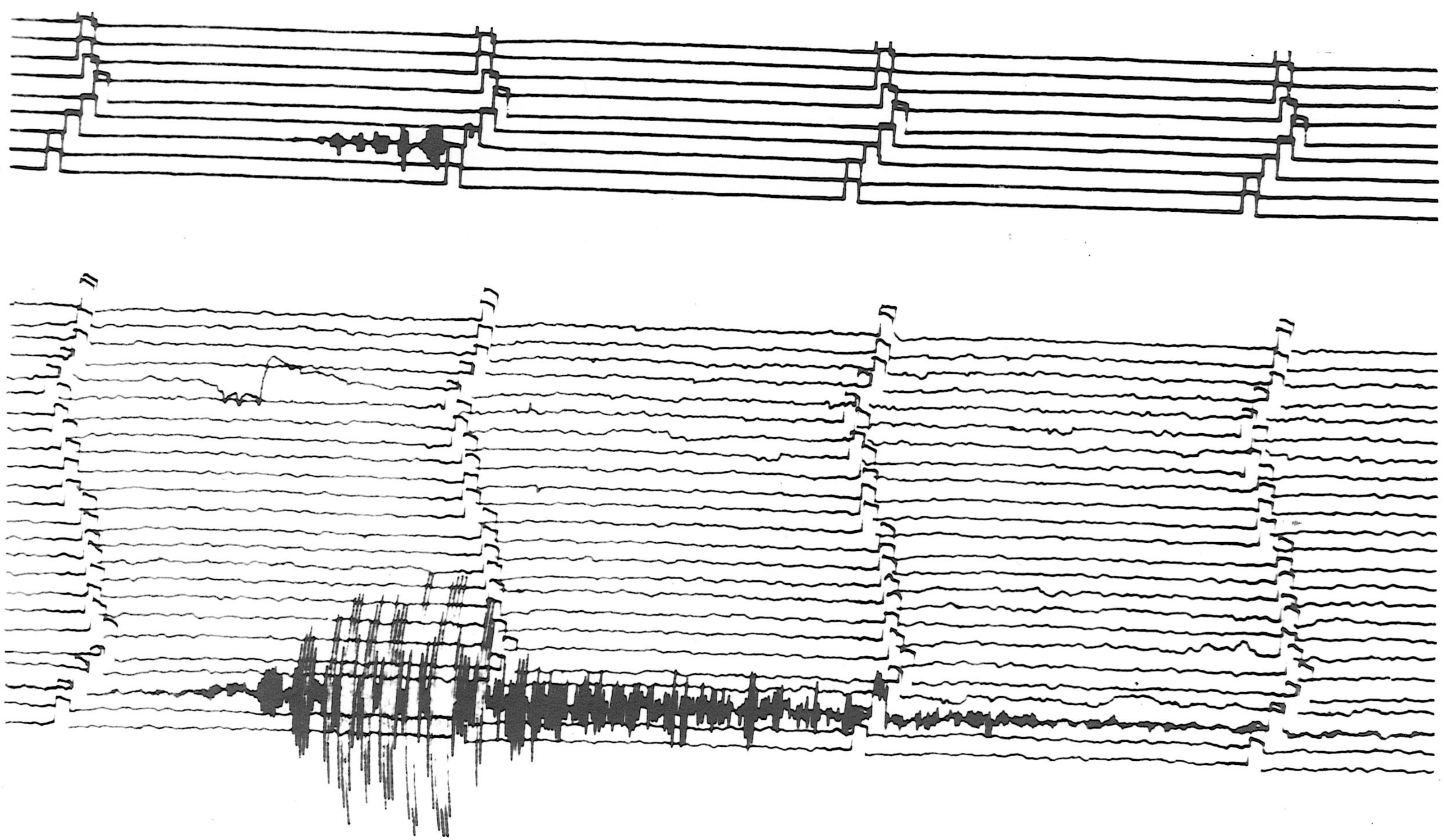

Fig. 10a. The T-phase of 24 December, 1963.
Top - TAU Newstead-Watt N-S, Low mag.
Bottom - TAU Newstead-Watt N-S, high mag.

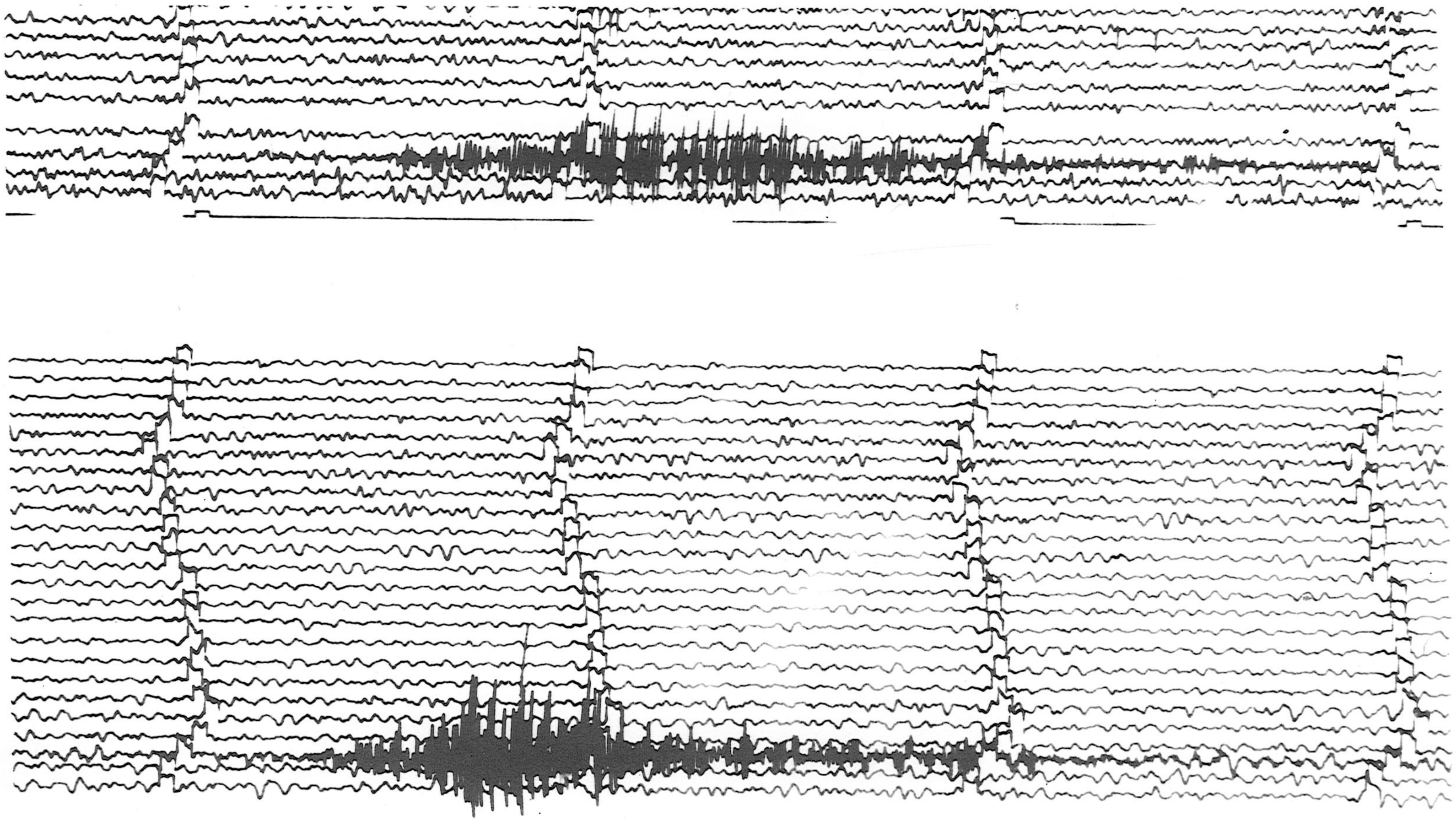

Fig.10b. The T-phase of 24 December, 1963.
Top - MOO, Willmore, Z.
Bottom - TAU, Hall-Sears, Z.

4.4 T Phase Propagation across Tasmania.

The apparent wave front velocity across Tasmania was determined for T phases recorded on at least three stations. Of 13 such cases, only two shocks (Table IV, Nos. 23, 24) gave reasonably clear T wave beginnings at all three stations. These had wave front velocities of 5.9 km/sec across Tasmania, which is the velocity of the crustal P wave.

Other apparent velocities measured on the seismic net were as low as 0.7 km/sec. The signals at the inland stations were too weak and emergent to be measured with any accuracy, although in some cases, the T phase at Hobart was very strong. The low velocities are therefore meaningless. Some energy may be transformed into S waves, velocity 3.5 km/sec, or surface waves, velocity 3 - 3.5 km/sec, but there is yet no evidence for these phases from the Tasmanian Seismic net.

The following are also sources of error in the velocity calculation. The wave front across Tasmania is assumed to be plane. Because of the uneven nature of the continental margin of south-east Tasmania, the wave front would be significantly distorted.

The T records from MOO (particularly), TRR and SAV are always much weaker and shorter than at

Hobart, which may result from strong attenuation of the T-phase, particularly at high frequencies, causing the beginning of the wave train to be read at Hobart, and a later part of it at the other stations. This would explain the variation of the measured velocities, the weaker T phases giving rise to lower velocities.

If different stations recorded different phases as the beginning of the train, (e.g. P at TAU, but S at TRR) erroneous velocities would result. But this situation is unlikely.

Different recording characteristics may be a reason for the poor quality of the inland station records. FNT and TAU signals are transmitted by short land line, but before TRR changed to land line transmission, MOO, TRR and SAV signals were all transmitted by a radio-telemeter system. Generally, TAU and FNT produce finer quality earthquake records, including T phases, than the other stations. If the radio links have a significant filtering effect or some other adverse effect on the high frequency component of the T wave signal, the measured arrivals of the radio-link stations will tend to be later than the true arrivals, giving rise to the observed slow velocity. Moorlands in particular is a poor T-wave recorder, since it is insensitive to frequencies greater than one cycle per sec.

Strong refraction of the energy at the continental margin is probable, but no reliable measurements of velocity and direction across Tasmania have yet been obtained to demonstrate this.

TABLE IV. MAJOR T PHASES RECORDED IN TASMANIA.

The epicentres are those given by U.S.C.G.S. Preliminary Determination of Epicentre Cards.

STATION

1	-	FNT
2	-	TAU
3	-	TRR
4	-	MOO
5	-	SAV
A	-	Canberra
B	-	Riverview
C	-	Toolangi (Melbourne)
D	-	Adelaide

No.	DATE	ORIGIN	STATIONS	REMARKS.
1.	31/12/57	45S 165.5E	1, B.	S.I.N.Z.
2.	24/ 5/60	44.5S 167.5E	1, B, C.	Felt Milford Sound. S.I.N.Z.
3.	25/ 5/60	44S 168E	1, A, B, C.	S.I.N.Z.
4.	13/12/60	52.1S 160.9E	1, 3, 4, A, B, C.	Two arrivals on FNT. Strong but emergent near Macquarie Island.
5.	14/12/60	51.9S 160.7E	3, 4, A.	Near Macquarie Island.

No.	DATE	ORIGIN	STATION	REMARKS.
6.	15/12/60		1, 3.	
7.	17/12/60		1, 3, 4.	Weak shock, poor records.
8.	26/12/60	49.4S 164.3E	1, 3, A, B, C.	Near Auckland Island.
9.	11/ 1/61	52.3S 160.3E	1, 3, A.	Near Macquarie Island.
10.	14/ 1/61	52.9S 160.8E	1, A.	Near Macquarie Island.
11.	27/ 2/61		1, 3, 4.	Very weak records.
12.	28/ 2/61		1, 3, A, B.	
13.	1/ 3/61		1, 3, 4, B.	Emergent.
14.	10/ 3/61	51.9S 161.6E	1, A.	Near Macquarie Island.
15.	18/ 3/61	49.9S 163.3E	1, 3, 4, A, B, D.	No timing on records Near Auckland Island.
16.	19/ 3/61		3, 4, A.	
17.	5/ 4/61	52.2S 160.0E	1, A.	Near Macquarie Island. 2 Arrivals.
18.	6/ 5/61	51.5S 161.3E	1, 3, 4, A.	Emergent.
19.	8/ 5/61		1, 3, 4.	Weak TRR and MOO P and T records.

No.	DATE	ORIGIN	STATIONS	REMARKS.
20.	12/ 6/61	49.6S 163.8E	1, B.	Near Auckland Island.
21.	4/ 7/61	43.2S 168.8E	1, 4, A, B.	S.I.N.Z.
22.	19/10/61	55.3S 146.4E	1, D.	Origin obscured by South Tasmania Ridge.
23.	5/11/61	49.4S 163.3E	1, 3, 4, 5, A, B.	Near Auckland Island velocity 5.9 km/sec across net.
24.	15/11/61		1, 3, 4, 5, A, B.	Velocity 5.9 km/sec across net, disregarding poor MOO record.
25.	5/12/61	50.8S 139.8E	1, 3, 5, D.	South West of Tasmania.
26.	19/ 1/62	49.8S 163.0E	1, A, B.	Near Auckland Island.
27.	17/ 4/62	42.6S 174.0E	1, A.	Weak T phase.
28.	10/ 5/62	41.8S 171.6E	1, A, B.	S.I.N.Z.
29.	17/ 5/62	41.9S 171.5E	1, 3, 4, A.	S.I.N.Z.
30.	16/ 7/62	52.1S 138.9E	2, D.	South west of Tasmania.

No.	DATE	ORIGIN	STATIONS	REMARKS.
31.	28/ 7/62		2, 3, 4, 5, A, B.	Emergent phase but 2 Arrivals.
32.	29/ 7/62	41.4S 173.2E	2.	Very faint.
33.	14/ 8/62	49.9S 163.0E	2, 3, 4, A, B.	2 T arrivals. Auckland Island.
34.	5/10/62		2, 3, 4, A, B.	Emergent.
35.	9/10/62		2, 3, 4, A, B.	Emergent. P absent or obscured.
36.	15/10/62	43.5S 169.8E	2, 4, A, B, C.	Emergent S.I.N.Z.
37.	13/ 1/63	49.7S 163.7E	2, 4, A, C.	Strong, Auckland Island.
38.	28/ 1/63	52.4S 159.6E	2, 3, 4, 5, A.	Emergent.
39.	30/ 5/63	54.2S 143.7E	2, D.	South Tasmania Ridge Magnitude $5\frac{1}{4} - 5\frac{1}{2}$ (Pal)
40.	10/ 6/63	55.4S 146.4E	2, C, D.	South Tasmania Ridge. Magnitude $6\frac{1}{2}$ (Pas).
41.	16/12/63	49.1S 127.1E	2, 3.	Indian-Aust. Ridge 5 degrees of path between 3000 and 4000 metre isobaths.
42.	24/12/63	53.0S 159.5E	2, 4.	Varying amplitude on Newstead-Watt. Horizontal N-S seismograph.

5. REGIONAL CRUSTAL STRUCTURE FROM SURFACE WAVE DISPERSION.

5.1 Surface Wave Dispersion.

The dispersion of surface waves in the period range 10 - 40 seconds is dependent on crustal structure. In oceanic regions, Rayleigh waves are mainly controlled by the water layer, and so are unsuitable for crustal study. Yamaguchi and Kizawa (1961) published a set of dispersion curves for the study of a wide range of crustal structures. The theoretical curves used in this study were taken from their paper.

The instruments used are the Fort Nelson Sprengnether seismograph and the Standard Set of long period Sprengnether seismographs installed at TAU by the U.S.C.G.S.. The arrival time of each is plotted against wave number , and by measuring the slope of the tangent along the curve, the wave period at any arrival time is obtained. Group velocity is obtained by dividing epicentral distance by travel time. The dispersion is plotted as a graph of group velocity against wave period. Epicentres and origin times are obtained from U.S.C.G.S. Epicentre cards.

5.2 Tasman Sea.

The amplitudes of Love waves on Tasmanian records of Melanesian earthquakes are often small, probably due to the non-uniformity of crustal structure across the Tasman Sea. However, the Love wave dispersion of two earthquakes was examined, and a value for the average crustal thickness of the Tasman Sea was obtained.

The dispersion curves assumed an upper layer with S velocity 3.66 km/sec, density 2.80 gm/cc, overlying a half space with S velocity 4.68 km/sec and density 3.00 gm/cc.

1. Loyalty Islands earthquake, , 22/12/1962, epicentre 22.0°S, 170.1°E. Path across the Norfolk Island Rise, Norfolk Island Trough, Lord Howe Rise, and Tasman Sea. (fig. 11).
   Crustal thickness from Love wave dispersion curve (fig. 12) approximately 12 km.
2. Kermadec Islands earthquake, 29/12/1962; epicentre 31.2°S 177.9°W.
   Path - Colville Ridge, across North Cape, of N.I.NZ. Lord Howe Rise, Tasman Sea. (fig. 11).
   Crustal Thickness from Love wave dispersion (fig. 12), 12 - 16 km, long coda due to low velocity ocean sediment which was not accounted for in theoretical model.

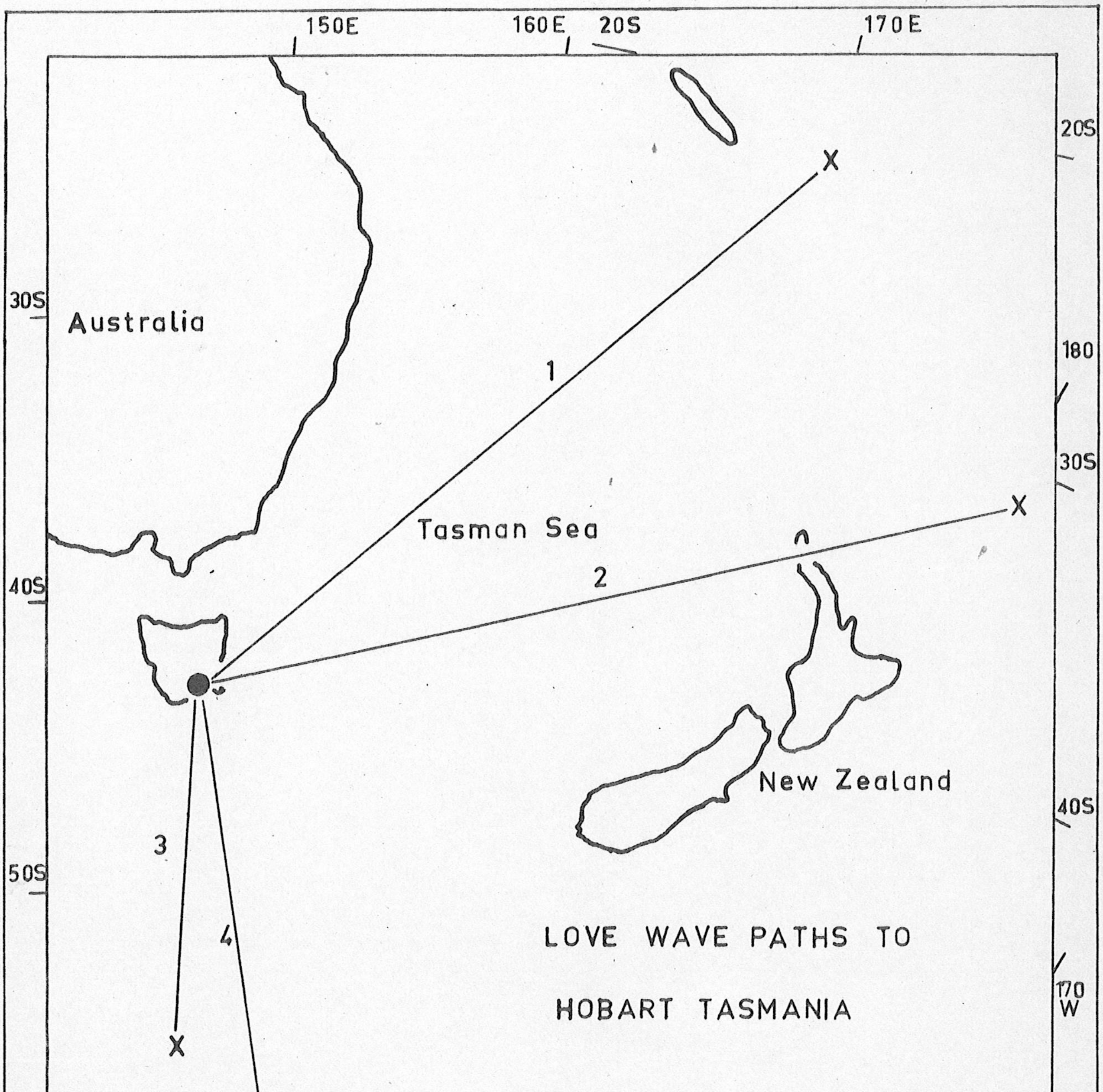

Fig. 11 Map showing Love wave paths of regional earthquakes used in crustal study.

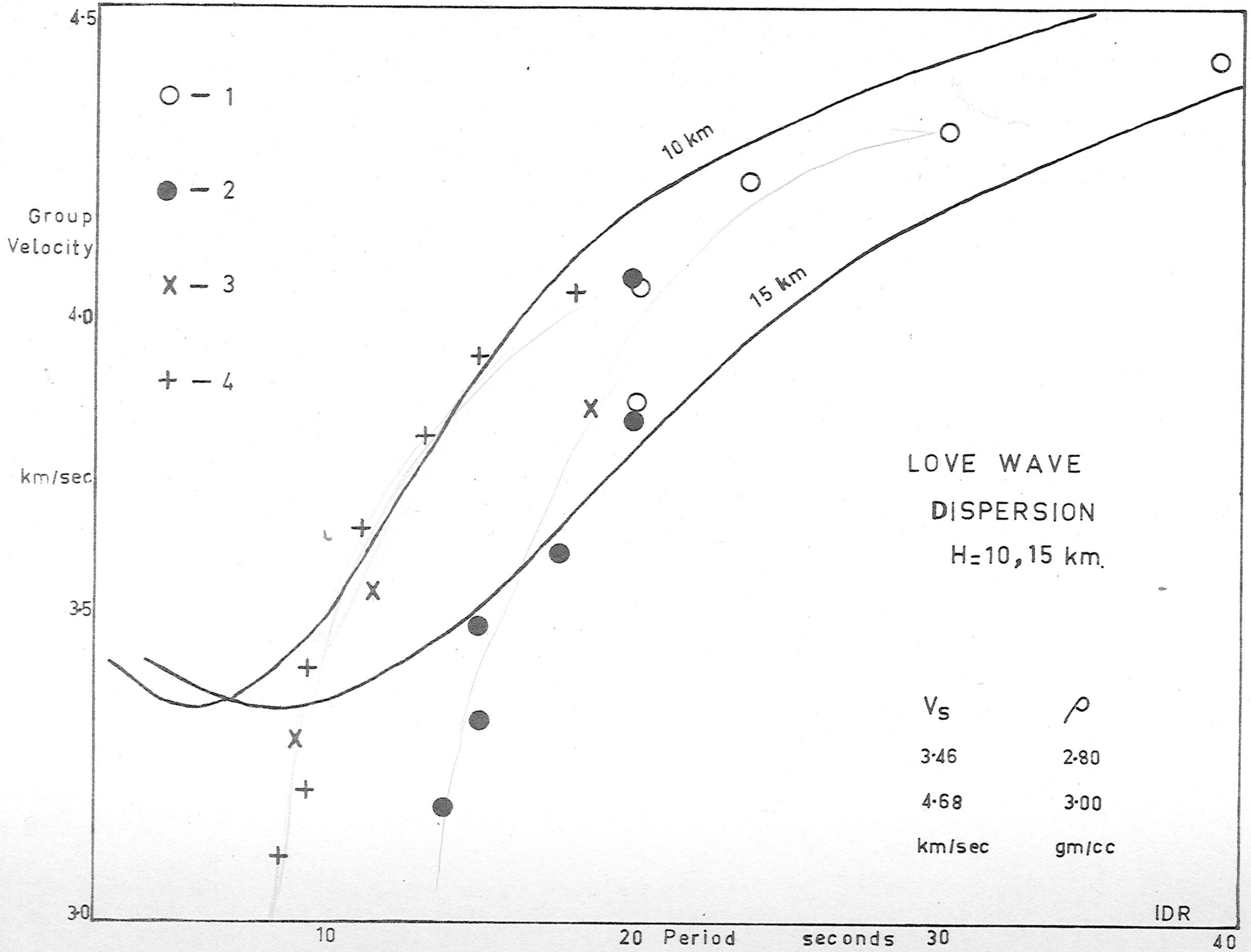

Fig.12. Love wave dispersion over Tasman Sea (1,2) and South Tasmania Ridge (3,4) paths. Numbers refer to wave paths shown in figure 11. (Theoretical curves after Yamaguchi and Kizawa, 1961).

From these two earthquakes, it is seen that the average crustal thickness of the Tasman Sea is approximately 12 km, thickening from the Loyalty Islands to New Zealand. From consideration of bathymetry, the crust would be thinnest below the Tasman Sea off southeast Australia, and thickening to the east towards New Zealand and north-east to Fiji and the New Hebrides.

5.3 <u>South Tasmania Ridge, Southern Ocean</u>.

An estimation of crustal thickness to the south of Tasmania was obtained Love wave dispersion of two earthquakes.

3. South Tasmania Ridge earthquake, 10/6/1963, epicentre 55.3°S 146.1°E.

   Path - due north, Indian-Antarctic Ridge, South Tasmania Ridge (fig. 11). Epicentral distance, 12°.

   Crustal thickness from Love wave dispersion (fig. 12), 10 - 14 km.

   Coda due to low velocity sediment which was not accounted for in theoretical model.

4. Earthquake, 1/12/1959. Epicentre 63°S 154°E.

   Path - Indian-Antarctic Ridge, South Tasmania Ridge, in Southern Ocean (fig. 11).

   Epicentral distance 20°. Recorded on original FNT Sprengnether.

   Crustal thickness from Love wave dispersion, 10 km (fig.12).

This earthquake, eight degrees further south than the previous, gives a smaller average crustal thickness, which shows that the crust is thinning gradually to the south through the South Tasmania Ridge. The region is not typically oceanic, for which the crust would be 5 km thick.

5.4 Australia.

Two earthquakes north of Australia gave surface wave paths across the Australia continent. But the paths also crossed anomalous regions north of Australia, and all that can be said from the surface wave dispersion is that it is consistant with an Australian crustal thickness of 30 - 35 km. The earthquakes were, - 21/12/1962, Java, 9.0°S 112.4°E, Rayleigh wave dispersion, and 2/1/1963, New Guinea 4.1°S 135.2°E, Love wave dispersion.

LIST OF REFERENCES.

BANKS, M. R., 1961. A comparison of Jurassic and Tertiary Trends in Tasmania. Dolerite, A Symposium, Univ. of Tas., 231 - 252.

BARTHOLOMEW, J., 1958, The Times Atlas of The World Vol. 1. The Times Publishing Co. Ltd. Printing House Square, London.

BULLARD, E. C., and D. T. GRIGGS, 1961. The Nature of the Mohorovicic Discontinuity. Geophys. J., 6, 118 - 123.

BURKE - GAFFNEY, T. N., 1954. The T-Phase from the New Zealand Region, Proc. Roy. Soc. N.S.W., 88, 50 - 54.

DOYLE, H. A., I. B. EVERINGHAM, and T. K. HOGAN, 1959. Seismic Recordings of Large Explosions in South-Eastern Australia. Aust. J. Phys., 12, 222 - 230.

EWING, W. M., W. S. JARDETZKY, and F. PRESS, 1957. Elastic Waves in Layered Media. McGraw-Hill Book Co. Inc.

EWING, M., and F. PRESS, 1956. Surface and Guided Waves, in Handbuch Der Physik, XLVII, 119 - 139. Springer-verlag, Berlin.

FLINN, E. A., 1960. Local Earthquake Location With An Electronic Computer. Bull. Seism. Soc. Am., 50, 467 - 470.

GREEN, R., 1962. Tasmanian Records of Earthquake T-Phases from New Zealand. N. Z. J. Geol. Geophys., 5, 322 - 330.

GUTENBERG, B., and C. F. RICHTER, 1942. Earthquake Magnitude, intensity, energy, and acceleration. Bull. Seism. Soc. Am., 32, 163 - 191.

JEFFRIES, H., and K. E. BULLEN, 1958. Seismological Tables. British Association for the Advancement of Science, Gray Milne Trust.

JONES, B. F., 1962. A Gravity Survey of the Great Lake Area, Tasmania. an Honours Thesis, University of Tasmania, Geology Department.

KANAI, K., 1957. The Requisite Conditions for The Predominant Vibration of Ground. Bull. Earthq. Res. Inst. (Tokyo), 35, 457 - 471.

KANAI, K., and T. TANAKA, 1961. On Microtremors, VIII. Bull. Earthq. Res. Inst. (Tokyo), 39, 97 - 114.

LINEHAN, D., 1940. Earthquakes in the West Indian Region. Trans. Amer. Geophys. Union., 19, 229 - 232.

NEWSTEAD, G. N., and A. BECK, 1953. Borehole Temperature Measuring Equipment and the Geothermal Flux in Tasmania. Aust. J. Phys., 6, 480 - 490.

NORTHROP, J., 1962. Evidence of Dispersion in Earthquake T-Phases. J. Geophys. Res. 67, 2823 - 2830.

OFFICER, C. B., 1958. Introduction to The Theory of Sound Transmission. McGraw-Hill Book Co. Inc.

RICHTER, C. F., 1958. Elementary Seismology. W. H. Freeman and Co., San Fransisco.

SPRY, A., and M. R. BANKS, 1962. Geology of Tasmania. J. Geol. Soc. Aust. Vol. 9. Part 2.

SYKES, L. R., 1963. Seismicity of the South Pacific Ocean. J. Geophys. Res. 68, 5999 - 6006.

UNIVERSITY OF TASMANIA. Geology Department Publication No. 84. "Tasmania University Seismic Net".

WOLLARD, G. P., and W. E. STRANGE, 1962. Gravity Anomalies and the Crust of the Earth. Amer. Geophys. Union, Geophys. Monograph No. 6, (The Crust of the Pacific Basin), 60 - 80.

YAMAGUCHI, R., and T. KIZAWA, 1961. Surface Waves and Layered Structures, (Part 2). Theoretical dispersion curves for sub-oceanic surface waves. Bull. Earthqu. Res. Inst. (Tokyo), 39, 669 - 809.

LOCALITY INDEX.

	Latitude	Longitude
Adelaide	34° 58'S	138° 42'E
Apsley	42° 25'S	147° 10'E
Auckland Island	51° 00'S	166° 00'E
Blackwood Creek	41° 44'S	146° 56'E
Bruny Island	43° 15'S	147° 20'E
Burnie	41° 03'S	145° 55'E
Canberra	35° 19'S	150° 00'E
Central Highlands	42° 00'S	146° 30'E
Cressy	41° 40'S	147° 07'E
Cygnet	43° 09'S	147° 06'E
Derwent Valley	42° 45'S	147° 00'E
Fiji	18° 00'S	178° 00'E
Fort Nelson	42° 56'S	147° 21'E
Gordon River	42° 40'S	146° 00'E
Gould's Country	41° 15'S	148° 00'E
Great Lake	41° 55'S	146° 45'E
Gretna	42° 40'S	146° 57'E
Hobart	42° 52'S	147° 20'E
Huon Valley	43° 05'S	147° 00'E
Indian-Antarctic Ridge	50°S	130°E
Java	8°S	114°E
Kermadec Island	30°S	179°W

		Latitude	Longitude
Lake Edgar		43° 00'S	146° 20'E
Lake Peddar		42° 56'S	146° 08'E
Lake St. Clair		42° 00'S	146° 10'E
Launceston		41° 24'S	147° 05'E
Loyalty Islands		21°S	167°E
Macquarie Harbour		42° 20'S	145° 20'E
Macquarie Island		54° 30'S	159° 00'E
Maydena		42° 45'S	146° 38'E
Melanesia	from	00S	140°E
	to	20°S	180°E
Milford Sound		44° 30'S	167° 00'E
Moorlands		42° 26'S	147° 11'E
Mount Barrow		41° 32'S	147° 23'E
Mount King William		42° 15'S	146° 10'E
Mount Wellington		42° 52'S	147° 15'E
New Guinea		5°S	140°E
New Hebrides		16°S	167°E
New Zealand	North Island	39°S	176°E
	South Island	44°S	171°E
North Cape		36°S	173°E
Orford		42° 33'S	147° 55'E
Port Davey		43° 20'S	145° 55'E
Queenstown		42° 05'S	145° 31'E

	Lattitude	Longitude
Riverview	33^{o} 50'S	151^{o} 09'E
Savannah	41^{o} 43'S	147^{o} 11'E
Serpentine River	42^{o} 45'S	146^{o} 00'E
Smithton	40^{o} 50'S	145^{o} 09'E
Solomon Islands	7^{o}S	147^{o}E
South Tasmania Ridge	47^{o}S	148^{o}E
Swansea	42^{o} 07'S	148^{o} 08'E
Tasmania University	42^{o} 55'S	147^{o} 19'E
Tasman Sea	40^{o}S	160^{o}E
Tarraleah	42^{o} 18'S	146^{o} 27'E
Tonga Islands	20^{o}S	175^{o}W
Toolangi (Melbourne)	37^{o} 34'S	145^{o} 291'E